Chapter 4: Sea Level Rise and Implications for Low Lying Islands, Coasts and Communities

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Table of Contents

Executive Summary

Sea Level Change

Sea level has risen in response to recent and past increases in temperature and will continue to do so over many centuries (*very high confidence*[1]). Geological evidence of sea level, and tide gauge and satellite observations provide evidence of this relationship over millennial, centennial, and decadal time scales. {4.2.2.1, 4.2.2.2, 4.2.2.6}

The geological record demonstrates that the polar ice sheets are highly sensitive to modest amounts of warming (*high confidence*). Peak global mean temperature during the Last Interglacial (130 to 115 thousand years ago) is estimated to be only 0.5°C–1.0°C warmer than pre-industrial, but sea level was 6–9 m higher (*medium confidence*). The Mid Pliocene Warm Period (~3 million years ago) was 1.9°C–3.6°C warmer than pre-industrial, and sea level was higher than during the LIG (*medium confidence*), but the maximum level remains deeply uncertain. The rates of past ice-sheet responses and sea level rise during these periods remain very uncertain. {4.2.2.1}

The rate of sea level rise is accelerating (*high confidence*). A combination of tide gauge records, satellite observations, and modelling shows that the average pace of 20th century sea level rise was slower than earlier estimates, implying that the pace of sea level rise in the last several decades has accelerated more than shown previously (*medium confidence*). {4.2.1.1}

Human activity was the predominant cause of global mean sea level rise since 1970 (*high confidence*). However, attribution of regional and local mean sea level change and individual events of extreme sea level is not yet possible. {4.2.2.6}

Rapid retreat and thinning of some Antarctic outlet glaciers is underway, pointing to the potential for dynamical ice processes to accelerate future sea level rise (*medium confidence*). Greenland ice loss will be dominated by ice-atmosphere interactions, rather than dynamic ice discharge to the ocean, limiting its potential effect on the rate and magnitude of sea level rise during the 21st century (*medium confidence*). {4.2.3.1}

Different modelling studies demonstrate that under high emissions scenarios, Antarctica will *likely*[2] contribute several tens of centimetres of sea level rise by the end of the century (*medium confidence*). Including Antarctica's dynamical contribution in projections of global mean sea level rise under RCP8.5 results in 0.89 m (0.66–1.13 m, *likely* range) for the period 2081–2100, and 1.06 m (0.82–1.33 m) in 2100, with respect to the reference period of 1986–2005. This magnitude and range is significantly greater than earlier assessments. {4.2.3.1}

Processes controlling the timing of future ice-shelf collapse and a possible Marine Ice Cliff Instability (MICI) make Antarctica's contribution to future sea level rise deeply uncertain for outcomes with probability outside the *likely* range (Cross Chapter Box 4). MICI in Antarctica has the potential to accelerate rates of sea level rise by the end of the 21st century and beyond, but the underlying physics in models lack explicit process-level details. {4.2.3.1}

Sea level rise at the end of the 21st century is strongly dependent on the climate scenario followed, especially in terms of Antarctica's contribution (*high confidence*). This points to the potential role of

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[1] FOOTNOTE: 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 1.8.3 and Table 1.2 for more details).

[2] FOOTNOTE: 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%, 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 1.8.3 and Table 1.2 for more details).

greenhouse gas mitigation in minimizing risk to low-lying coastlines and islands. For the first half of the 21st century differences among the scenarios are small. {4.2.1.2}

Extreme sea level (ESL) events associated with disastrous flooding which are historically rare, will become common by 2100 under all emission scenarios (*high confidence*). The emission scenario determines the rate at which ESLs become more frequent and the height of these ESLs. Sea level rise will amplify the height and frequency of ESLs, whether or not coastal storms intensify. Under all scenarios, ESLs that are historically rare (e.g., 0.01 annual probability), are projected to become annual events by 2100 at many low-lying coastal areas (*high confidence*). Under RCP8.5, several regions including small islands in the western Pacific and some megacities will experience such annual ESLs by 2050. The sensitivity of coastal areas to projected increased height of ESLs by 2100 differs regionally and between emission scenarios. {4.2.3.4}

Subsidence and changes in ocean wave characteristics are important factors in estimating future changes in relative sea level (RSL) and ESL (*high confidence*). In some regions, changes in wave height and period currently have a larger effect on coastal flooding than RSL change (*medium confidence*). Subsidence caused by human interventions is currently the most important cause of RSL change in many delta regions. While the relative importance of sea level-rise will increase over time, this implies that a consideration of local processes is critical for projections of sea level impacts at local scales. {4.2.1.6, 4.2.2.5}

Exposure and Vulnerability

Anthropogenic non-climatic drivers played the major role in increasing exposure and vulnerability to sea level rise and extreme sea levels worldwide over the course of the last century; and they will continue to do so, in the absence of adequate, proactive adaptation, through to the mid-21st century (*high confidence*). This has been confirmed by recent literature, which has made progress in more systematic and comprehensive assessment of all dimensions of SLR-related coastal risk (i.e., hazard, exposure and vulnerability). This includes improved projections of sea level rise and extreme sea levels, and their associated impacts globally and regionally (4.2.1.2 and 4.2.3.4). Exposure assessments have improved by considering spatial and temporal dynamics of current and future exposure. Vulnerability assessments have improved by coupling social and ecological dynamics, assessing multiple hazards, employing vulnerability functions and thresholds, and using improved data sources (*medium evidence, high agreement).* {4.3.1}

Coastal nations, and low-lying coasts in particular, are exposed and vulnerable to SLR and extreme sea levels, but the degree of SLR-related coastal risk is determined by context-specific circumstances (*very high confidence*). Although attribution of observed impacts primarily to local SLR and extreme sea levels or to other anthropogenic drivers remains challenging, there is compelling evidence that non-climatic factors significantly influence local risk, impacts, response options and barriers for adapting to climate change. This has important implications for decision-making because effective actions can be undertaken in the short-term to target local drivers of exposure and vulnerability, notwithstanding uncertainty about local climate change impacts in coming decades and beyond. In addition, understanding about the diversity and interactions of the anthropogenic drivers of exposure and vulnerability is improving. {4.3.2.1}

Impacts

Sea level rise and extreme sea levels pose major threats to coastal social-ecological systems and associated communities worldwide, and to low-lying coastal cities, islands, deltas, and the Arctic in particular (*medium evidence, high agreement*). These threats will manifest in both direct (e.g., coastal flooding, salinization, etc.) and indirect (e.g., economic, social, and ecological repercussions from flooding, etc.) ways. {4.3.3.1}

Climate change impacts are being observed in a growing number of places, including impacts on biodiversity and ecosystem services, coastal infrastructure, community livelihoods, and cultural and aesthetic values (*medium evidence, medium agreement*). However, attribution of impacts to sea level rise per se is difficult due to the influence of local processes unrelated to climate (e.g., subsidence, coastal development, sediment transport). Observed and potential impacts arise from the nature of in situ ecosystems (coral reefs, wetlands and saltmarshes, beach-dune systems, etc.) and anthropogenic dynamics (e.g., coastal urbanization) in addition to sea level rise and extreme sea levels. A combination of drivers has affected these components in recent decades, leading, among other things, to degradation of coastal and marine ecosystems, coastal squeeze, sediment starvation upstream and human-induced subsidence (*medium evidence, medium agreement*). {4.3.3.3}

Responses

A variety of coastal adaptation responses, including measures to protect, advance, accommodate and retreat, are available and being applied around the world to cope with climate variability and reduce coastal hazard risk. Responses seldom explicitly target sea level rise (*high evidence, medium confidence*). Recent literature recognizes that adaptation efforts are underway at various spatial scales, in various geographical and territorial contexts, and through a wide diversity of interventions, including ecosystem-based and community-based approaches, hard and soft coastal defences, climate risk management plans, and human mobility that includes planned relocation of people and activities {4.4.2}.

Hard and soft engineering-based coastal protection and advance measures are especially widespread in low-lying urban and densely populated coastal areas, and will expand in the future because these measures are highly effective and cost-efficient for urban areas in the 21st century even under high SLR (*medium agreement, medium evidence*). There is *medium confidence* that well-designed and maintained hard and soft protection provides predictable levels of safety and there is *medium confidence* regarding design considerations for hard protection and sediment-based measures. Design considerations for coastal protection measures can account for future sea level rise and are increasingly implemented (*limited evidence, high agreement*). {4.3.3.3.1, 4.4.3.1, 4.4.4}

Ecosystem-based and hybrid (combinations of natural and built infrastructure) solutions are gaining traction worldwide and progress has been made to demonstrate their effectiveness, identify co-benefits, and quantify costs and benefits (*medium evidence, high agreement)*. Vegetation (marshes/mangroves, seagrasses/kelp, mussel beds) and reefs (coral/oyster) provide protection and risk reduction benefits to those living in nearby coastal locations (*medium evidence, high agreement)*. There is *medium evidence* that ecosystem-based measures bring substantial economic benefits, but *low agreement* regarding the actual size of the benefits. However, ecosystem-based measures provide multiple additional co-benefits (*high confidence*). Due to their space requirements, ecosystem-based measures play a smaller role in densely populated urban areas. There is *medium evidence and low agreement* regarding design considerations for ecosystem-based measures. {4.4.2, 4.4.3.2, 4.4.4.2, 4.4.4.5}

Despite deep uncertainty about long term future mean and extreme sea levels, adaptation can be progressed in the short-term by applying decision-analytical methods (*medium evidence, high agreement*). These methods range from consideration of high-level adaptation pathways approaches that can be applied in diverse contexts, to technical and costly methods of robust and flexible decision making that can be applied to assess specific, large scale investment decisions. More integrated and systematic collaboration between researchers on sea level rise, decision sciences and decision makers has the potential to improve the methods and decisions made today in the face of anticipated sea level rise and extreme sea levels (*limited evidence, medium agreement*). {4.4.5.2, 4.4.5.3}

Community-based approaches are increasingly used by people living in low-lying coastal areas to adapt to climate change, especially in developing countries (*medium evidence, high agreement*). Community-based adaptation aims to involve local people directly in understanding and addressing the climate change risks they face, including SLR-related risk. In particular, this approach seeks to reduce local-level vulnerability and build resilience, especially of those most at risk. {4.4.5.4}

Community-based adaptation is more likely to be effective when it is an integral part of more general community development efforts (*limited evidence, high agreement*). The drivers of poverty, inequity and political marginalization shape vulnerability and SLR-related coastal hazard risk and they are not readily addressed by ad hoc community-based adaptation projects. Hence the need to integrate such efforts into development initiatives more generally. {4.4.5.4}

Participatory approaches, community visioning and consensus building are an essential part of community-based adaptation but realizing the potential of such approaches is difficult to achieve in practice (*medium evidence, high agreement*). Authentic and meaningful engagement by community members in community-based adaptation is recognized as being essential. However, unless extreme events have been experienced, or the prospect of SLR impacts is readily apparent, pressing immediate needs tend to dampen community efforts to take action to address seemingly distant and uncertain issues like SLR. Moreover, in many settings, powerful interests prevail and vulnerable groups are marginalized in local planning and decision-making. {4.4.5.4}

Adaptation pathways have emerged as an important way to frame thinking about responses to climate change, and rising sea levels in particular, because this approach recognizes and enables sequenced long-term decision-making in the face of dynamic and deeply uncertain SLR-related coastal risk (*medium evidence, high agreement*). {4.4.5.3, 4.4.5.4}

The imperative to integrate climate change mitigation and adaptation measures that enable Climate Resilient Development Pathways has been brought to the fore by this Chapter. If greenhouse gas emissions continue along current trajectories, SLR at the end of the 21st century will have devastating impacts on low-lying coasts and islands. Transformative mitigation and adaptation responses need to be institutionalized to enable climate resilient development pathways.

4.1. Purpose, Scope, and Structure of the Chapter

The objective of this chapter is to assess literature on sea level rise and its implications for low-lying islands, coasts and communities published since the Fifth Assessment Report (AR5). The chapter comprehensively evaluates the current and future state of low-lying islands, coasts and communities in the context of sea level rise, changing characteristics of extremes of coastal high water, and related consequences of climate change. Owing to the nature and magnitude of the expected effects of sea level rise on human coastal communities and the interlinked nature of exposure and vulnerability of human communities and coastal ecosystems, the chapter's focus is on socio-ecological systems and not on marine and cryosphere ecosystems. We also take cognizance of the IPCC Special Report on Global Warming of 1.5°C (SR1.5) as an additional point of departure.

For assessing coastal exposure, vulnerability, and risk to ecosystems, species, and human systems, groups, and individuals, this chapter adopts the risk framework developed in the Special Report on Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation (SREX). This framework was applied extensively in the AR5 and subsequent literature. This chapter also assesses pathways to resilience and sustainable development along the coast in the specific context of climate change and sea level rise. In each aspect, we place special emphasis on Small Island Developing States (SIDS), deltas, and other low-lying coastal areas, including coastal cities.

4.1.1 Themes of this Chapter

Section 4.2 establishes the physical setting relevant to coastal hazards including the numerous processes contributing to sea level rise globally, regionally, and locally, as well as dynamical coastal morphology. This section also assesses paleo-climate evidence, direct observations of sea level change and its acceleration, and attribution of sea level rise, and assesses previous projections in light of new literature with a particular focus on ice-sheet dynamics and uncertainties. Section 4.3 assesses socio-economic and demographic drivers of change with an emphasis on how human and societal factors, such as urbanization, interact with the ever changing social-ecological coastal setting to determine the dynamic pattern of coastal exposure and vulnerability for ecological and human systems. These socio-economic and demographic drivers of change interact with the changing hazards evaluated in Section 4.2 to determine risk and impacts (the manifestation of risk). Cultural, institutional, and ethical dimensions are assessed insofar as they both influence and are influenced by risk and impacts. Governance and the interactions across scales of governance, such as individuals' interactions with local, regional, national and transnational institutions, are an especially important focus. Also important are the ways that risk perception affects responses to risk. The Reasons for Concern and the associated Burning Embers diagram, a global aggregation of risk established in the Third Assessment Report (AR3), are updated in the context of sea level rise and coastal risk. Section 4.4 assesses literature on responses, particularly options and development pathways facilitating planning and implementing adaptation, building resilience, and facilitating transformation. The section concludes with an assessment of pathways to resilience and sustainable development, including measures, safety margins, barriers and enablers of response.

Figure 4.1 illustrates the interconnection of this chapter's themes, including drivers of sea level change and extreme sea level hazards (Section 4.2), drivers of exposure, vulnerability, and impacts and risk related to sea level change (Section 4.3), and climate resilient development, with a focus on responses to sea level rise (Section 4.4). Our approach is to anchor this discussion throughout the chapter with specific examples of coastal risk and individual and institutional responses to risk and its manifestation (impacts) as they evolved over time in specific, placed-based contexts and events.

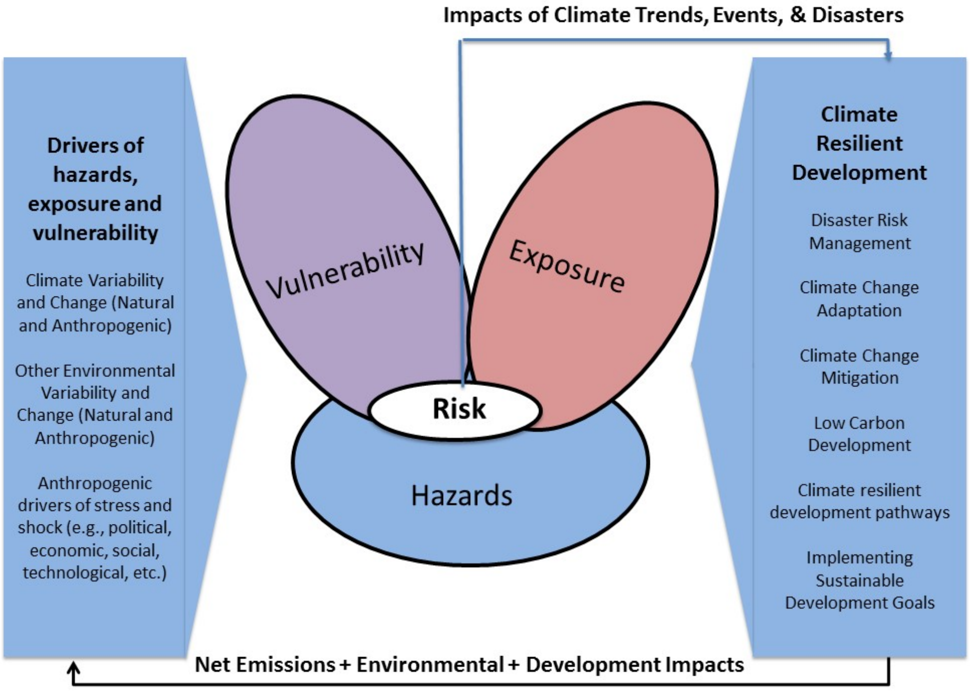


Figure 4.1: Schematic illustration of the interconnection of Chapter 4 themes, including drivers of sea level change and extreme sea level hazards (Section 4.2), drivers of exposure, vulnerability and impacts and risk related to sea level change (Section 4.3), and climate resilient development, with a focus on responses to sea level rise (Section 4.4).

4.1.2 Advances in this Chapter Beyond AR5 and SR1.5

[PLACEHOLDER FOR SECOND ORDER DRAFT: material from SR1.5 to be added]

4.1.2.1 Coastal Hazards

AR5 assessed past sea level change based on the instrumental and geological record and projected a *likely* range for global mean sea level rise (0.28–0.98 m by 2100 compared to 1986–2005; (e.g., Church et al., 2013). Importantly, AR5 also assessed regional sea level changes, and changes in extremes of high water, especially as they relate to flooding. AR5 presented IPCC's first quantification of the dynamical contribution of the Antarctic ice sheet to sea level rise by 2100 but cautioned that, 'Based on current understanding, only the collapse of marine-based sectors of the Antarctic ice sheet, if initiated, could cause global mean sea level to rise substantially above the *likely* range during the 21st century. This potential additional contribution cannot be precisely quantified but there is *medium confidence* that it would not exceed several tenths of a meter of sea level rise during the 21st century.' An important advance in this chapter is our assessment of the literature since AR5 relevant to projecting the Antarctic ice sheet contribution (Section 4.2.3). We further address the tail or high end of the probability distribution of sea level rise (Section 4.2.3) while also elaborating on the deep uncertainty (i.e., ambiguous or difficult-to-quantify probabilities) that inhibits full characterization of the tail (Cross Chapter Box 4; Section 4.2.3.4; Oppenheimer et al., 2016; Bakker, 2017).

Literature since AR5 assessed in this chapter leads us to reevaluate AR5 projections of changes in frequency of regional extremes of high water associated with coastal storms and flood events (Section 4.2.3.4). AR5 projections of regional sea level did not include all components, such as tectonics or subsidence associated with groundwater and hydrocarbon withdrawal, and their absence increases uncertainty in projection of extremes. AR5 projections of regional extremes of high water were also limited by uncertainty in projecting characteristics of tropical and extratropical cyclones. Several of these uncertainties remain and limit confidence in updated projections of extremes for some regions and time periods.

This chapter also updates previous assessments of coastal high water and extreme precipitation occurring simultaneously, subsidence, and waves in light of recent advances. Taken together, improved understanding of coastal hazards gives us higher confidence in evaluating risk.

4.1.2.2 Exposure, Vulnerability, Impacts and Risks

AR5 addressed coastal vulnerability and exposure to natural and human systems at many regional settings in its comprehensive assessment of risk, finding with *very high confidence* that "Coastal systems and low-lying areas will increasingly experience adverse impacts such as submergence, coastal flooding, and coastal erosion due to the combined effect of relative sea level rise and extreme sea level." AR5 also updated the Reasons for Concern framework and the Burning Embers representation of risk to incorporate additional elements: risk for marine species impacted by ocean acidification only, by the combined effect of acidification and warming, and risk for coastal human and natural systems impacted by sea level rise (Wong et al., 2014). Ability to perform a detailed assessment of future risk was constrained by the uncertainty in regional projections of sea level change, extreme sea level, and changes in storm frequency and intensity, noted in the previous subsection. Improved hazard assessment and a significant extension of the literature relevant to more spatially and temporally explicit current as well as future coastal exposure and vulnerability, allows increased confidence in the assessment of coastal risk in SROCC.

4.1.2.3 Responses

This chapter describes the variety of observed and available responses, where and to whom they have been applied currently, costs, benefits and co-benefits of response options, frameworks for appraising and selecting appropriate options, as well as limits and barriers to their implementation.

We evaluate the implications of three key characteristics of future climate and sea level for responses. These aspects are 1) growing uncertainty in climate change arising from the increasing differences over this century between the Representative Concentration Pathways (RCPs) beyond 2050; 2) growing uncertainty in global and regional sea level rise, due to uncertainty in the dynamical contribution from the Antarctic ice sheet, especially in the latter half of this century; 3) a resulting increase over time in uncertainty of estimated return periods for sea level extremes associated with storms and coastal flooding. These aspects lead to projections of progressively greater uncertainty in risk with time and bear implications for effective response strategies (Section 4.4).

4.2 Physical Basis for Sea Level Change and Associated Hazards

As a consequence of natural and anthropogenic changes in the climate system, sea level changes are occurring on temporal and spatial scales that can cause increased levels of risk for coastal communities, cities, and low-lying islands. On a global scale, sea level rise is caused by volume changes of ocean water, or by mass changes caused by loss of land ice or changes in terrestrial water reservoirs. Mass changes lead to distinct spatial patterns of regional sea level rise, often called fingerprints. These fingerprints are caused by gravitational and Earth rotation changes as masses of ice and water are redistributed on the Earth's surface. Here regional sea level refers to spatial scales of around 100 km, while local sea level refers to spatial scales smaller than 10 km. In addition to gravitational and rotational effects, the solid Earth responds to changes in both long processes—including tectonics, mantle dynamics, post-glacial rebound—and short-term redistribution of water, ice and sediments, both natural and/or anthropogenic in origin. This causes vertical land motion and ocean surface changes at coastlines, and hence a relative sea level change (RSLC), defined as the difference in elevation between the land and sea surface at a specific time and location (Farrell and Clark, 1976). In most places around the world, current annual mean rates of regional and relative sea level changes are typically in the order of a few mm yr^{-1} (see Figure 4.4). Additional risk associated with changing sea level is related to individual events, superposed on the background of gradual change. As a result, the gradual changes in time and space may be assessed together with processes that lead to flooding events but have a broad spectrum of variability. These include storms, surges, waves, and tides or a compounded combination of these processes that can lead to extreme sea level events. In this section, newly emerging understanding of these different episodic and long-term aspects of sea level change are assessed, within a context of sea level changes measured directly over the last century, and those inferred for longer geological timescales. This longer-term perspective is important for contextualizing future projections of sea level and providing guidance for processes-based models of the individual components of sea level rise, including the polar ice sheets.

4.2.1 Processes of Sea Level Change

Sea level changes have been discussed throughout the various IPCC assessment reports as sea level rise is a key feature of climate change. Beginning with the First Assessment Report (AR1), it was realized that thermal expansion of the oceans and the mass loss from glaciers were important drivers of the observed changes. In addition, the slow response time of the cryosphere and ocean, in combination with ongoing warming in the near future, implied the potential for substantial future sea level rise, even if greenhouse gas (GHG) emissions are reduced with respect to current rates (Warrick and Oerlemans, 1990). In the early 1990s, observed changes in the polar ice sheets covering Greenland and Antarctica were small, and the general understanding, based in part on numerical ice sheet models (Huybrechts, 1994) estimating global ice volume changes, was that they would not provide a major contribution to future sea level on decadal or even century timescales. In fact, it was assumed that increased snow accumulation rates in Antarctica in response to a warming polar atmosphere would contribute to a small net drop in sea level. Complex interactions between the oceans and ice sheets were not yet recognized as important drivers of processes that can lead to rapid dynamical changes in the ice sheets. Understanding of ice calving processes and glacial hydrological processes was also limited. The view on the potential role of ice sheets in future sea level rise changed by the time of AR4 (Lemke et al., 2007), following the first convincing signs of increased ablation rates in Greenland and increased rates of ice discharge into the ocean around Antarctica. By then, projections of future sea level were presented with the caveat that dynamical ice-sheet processes were not accounted for as our physical understanding of these processes was still too rudimentary (Bindoff et al., 2007). This implies that processes related to changes in the atmospheric conditions were captured, but the adjustment of the ice flow to the changed environmental conditions were not. In AR5 (Church et al., 2013) a first attempt was made to quantify the dynamic contribution of the ice sheets, although still with limited physics included and mainly relying on an extrapolation of existing observations. A second major point of progress in the AR5 report was the improved insight into local and regional patterns of sea level change. These two points provide the basis for this chapter, where we focus on sea level changes around coastlines and low-lying islands, rather than global mean sea level rise. Here we explain the mechanism driving past and contemporary sea level changes and episodic extremes of sea level and assess confidence in regional projections of future sea level over the 21st century and beyond. While the emphasis is on progress since AR5 published in 2013, we note that new climate model intercomparison results (CMIP6), providing updated guidance on specific components of sea level rise including ocean thermal expansion and circulation changes, ocean temperatures in contact with ice sheets, and evolving atmospheric temperatures above glaciers and ice sheets, are not yet available.

4.2.1.1 Global Mean Sea Level and Relative Sea Level

Changes in the volume of ocean water control global mean sea level (GMSL) on timescales ranging from decades to centuries (Church et al., 2013). Ocean volume is a function of both the mass of sea water and its density (or its inverse, specific volume). The total mass of the ocean changes as the partitioning of fresh water reservoirs (ground water, lakes and land water storage, glaciers, and the ice sheets on Greenland and Antarctica) move between the land and the ocean. Tectonic and other dynamic Earth processes including dynamic topography of the Earth's surface, and glacial isostatic adjustment (GIA) caused by past changes in ice sheets, also impact GMSL, through their effect on the geography and mean depth of the underlying sea floor, and the Earth's gravitational field (Tamisiea et al., 2010).

Since the publication of AR5, a combination of approaches using information from ancient (paleo) sea level records, the global network of tide gauges, and satellite data have substantially advanced our understanding of sea level change over the past century and beyond. Over the last century, about 50% of this change in sea level was caused by thermal expansion of the global ocean, however the addition of ocean mass from the loss of land ice has begun to outpace thermal expansion as the dominant contributor since around 2005 (Table 4.1). Since 1993, a combination of tide gauge and satellite-based estimates consistently indicate a sharp increase in the rate of GMSL rise, with the rate of SLR accelerating within the satellite era (Nerem, 2018).

Sea level does not rise uniformly, but exhibits substantial regional variability at decadal to multi-decadal time scales. Changing winds, air-sea heat and freshwater fluxes, and the addition of riverine and glacial meltwater alter ocean currents, which lead to regional and local changes in sea level. Contemporary and past changes in land ice cover perturb the gravitational field of the Earth, deform the Earth's crust and change the orientation and rate of the Earth's rotation. In turn, these processes affect sea level regionally, and can make it deviate substantially from the global mean sea level (Mitrovica et al., 2011).

Global mean and regional sea level changes are useful concepts for considering general trends in sea level, however, it is relative sea level (RSL) that directly impacts coastal communities, cities and low-lying islands. An analysis of RSL may take into account regional to local changes in the Earth's geoid as water and ice mass move over the Earth's surface, vertical motions of the sea floor and coastal regions, and in some places subsidence due to changes in the delivery and compaction of sediment and landfill, and extraction of subsurface freshwater and hydrocarbons. In sum, changes in RSL are caused by multiple, interacting, and sometimes compounding factors like storms and high tides. As a result, reliable projections of future RSL at specific time projections and locations remain difficult to make.

4.2.1.2 Ice Sheets and Ice Shelves

The vast majority of water in the cryosphere is stored in the ice sheets of Antarctica and Greenland (see Chapter 3). Nevertheless, glaciers have contributed more to GMSL rise over the last century, due to their faster response time and location in relatively warmer climate zones (e.g., Gregory et al., 2013b). The large ice sheets on Greenland and Antarctica are not expected to disappear on centennial timescales, but because of their volume, the loss of even a small fraction of their mass could begin to dominate sea level rise. It is therefore of utmost importance to understand how ice sheets' mass can change with time. Figure 4.2 illustrates the most important processes that drive mass change of an ice sheet. The total mass of an ice sheet is controlled by the surface mass balance (SMB), the sum of ablation and accumulation controlled by atmospheric processes. Furthermore, ice sheets lose mass through contact with warm ocean water below the ice shelves and by iceberg discharge at the ocean margin. Changes in the SMB, discharge, and melting forced by the ocean will lead to a dynamical adjustment of the ice sheet. Ice sheets drive changes in sea level mainly through the loss or gain of land ice above flotation, which is the ice thickness above local sea level, corrected for density difference between water and ice. At present Greenland is contributing more to sea level change than Antarctica, but significant parts of Antarctica are resting on bedrock below sea level, which has a large potential to contribute to sea level via a dynamical response to ocean melt, and a possible marine ice sheet instability (MISI, see Section 4.2.3.1 and Chapter 3).

While changes in floating ice shelves do not contribute directly to sea level change, they play an important role in the dynamics of ice sheets. They gain mass through the inflow of ice from the ice sheet. The surface balance of the ice shelves may be either positive or negative. At present, accumulation is larger than ablation in most areas, but changes in ablation are usually more important in a changing climate than the changes in accumulation. If the ablation is substantial, the shelves are not only losing mass, but penetration and movement of the surface meltwater can deepen crevasses at the surface, and cause stresses that can lead to hydrofracturing and ice shelf collapse. This has been witnessed on the Larsen A, Larsen B, and Wilkens ice shelves on Antarctica (Scambos et al., 2000; Banwell et al., 2013; Macayeal and Sergienko, 2013; Kuipers Munneke et al., 2014). In addition, melt on the bottom side of shelves is controlled by the circulation, temperature, and salinity of the water. These water properties determine the magnitude and pattern of melt, and refreezing of ocean water. As the pressure melting point of ice increases with water depth, melt tends to dominate, especially near deep grounding lines, the transition where seaward-flowing grounded ice begins to float (Figure 4.3), and where marine-terminating ice margins begin to float. Iceberg formation at either the grounding lines or ice-shelf fronts is governed by complex ice-mechanical processes, the internal strength of the ice, and interaction with ocean waves and tides (Benn et al., 2007; Bassis, 2011).

Ice shelves are critically important, because they have a buttressing effect on the ice sheet behind the shelf. Their thinning or disappearance leads to an acceleration of the grounded ice upstream (Weertman, 1974; Thomas, 1979; Scambos et al., 2004; Schoof, 2007b). Subsurface ocean and surface air warming can lead to the complete disintegration of ice shelves, triggering MISI. It is estimated that the potential amount of ice discharge in response to melting of ice shelves buttressing marine based ice is equivalent to 3.5 meter for West-Antarctica alone (Bamber et al., 2009). The process of marine ice sheet stability is particularly important in West-Antarctica where large parts of the ice sheet are based on bedrock below sea level.

In addition, it is postulated that disappearance of ice shelves allows formation of ice cliffs, which may be inherently unstable if sufficiently tall (Bassis and Walker, 2012). Their collapse can lead to ice sheet retreat, through a process called marine ice cliff instability (MICI, see section 4.2.3.1.2 and Box 4.1), that may contribute to significant mass loss in both West and East Antarctica (Pollard et al., 2015; DeConto and Pollard, 2016). However, few direct observations are available to constrain the importance of ice-cliff failure, and those observations that do exist are in relatively narrow outlet glaciers on Greenland, which might not be appropriate analogues for the larger spatial scales of many Antarctic glaciers.

Our understanding of ice sheets has progressed substantially since the AR5, although deep uncertainty (Cross Chapter Box 4) remains with regard to their potential contribution to future sea level rise after the first half of the 21st century. This is particularly true for Antarctica.

4.2.1.3 Glaciers

Glaciers and ice caps not associated with Greenland and Antarctica also contribute to sea level change (Figure 4.2). They gain mass by accumulation (mainly snowfall) and lose mass by ablation (mainly melt) or calving in lakes or the ocean. Because of their, on average high accumulation and ablation rates compared to the ice sheets, they are sensitive indicators of climate change and respond fast to change in the climate system, with a response time scale in the order of decades. As a consequence, glaciers added more mass to the ocean than the Greenland and Antarctic ice sheets during the past century (e.g., Gregory et al., 2013b). However, the volume of ice in this land ice reservoir is small by comparison, equivalent to only ~0.4 m sea level rise if all the world's glaciers were lost (Arendt et al., 2012). In some areas, the loss of glaciers has been interrupted by periods of increased precipitation or regional cooling (Mackintosh et al., 2017), but on longer time scales, the impact of global temperature tends to dominate, contributing to ongoing glacial melt.

4.2.1.4 Ocean Processes

Warmer ocean water has a lower density and therefore a larger volume per unit mass, and consequently this leads to higher sea level even when ocean mass is constant. Over at least the last 1500 years sea level variability was tightly coupled to global mean temperatures (Kopp et al., 2016), partly due to ice mass loss and partly by thermal expansion of ocean water. In the past century, thermal expansion was the single greatest contributor to GMSL rise, although increased ocean mass mainly through ice loss is now (since ~2005; (Shepherd et al., 2012; Church et al., 2013; A. Cazenave, 2018)), the dominant contributor.

More than 90% of the increase in energy in the climate system was stored in the ocean over the last decades, implying that sea level and climate change are intimately related and hence that thermal expansion provides insight in our understanding of the climate system and climate sensitivity. Findings from these two fields are consistent (Otto et al., 2013). In addition to steric expansion changes in the ocean dynamics and salinity also play a role in regional sea level changes.

4.2.1.5 Terrestrial Reservoirs

Finally, global sea level is affected by changes in terrestrial reservoirs of liquid water. Withdrawal of groundwater and storage of fresh water behind man-made dam construction contributes to sea level change. In the earlier parts of the 20th century the terrestrial contribution was dominated by the storage component, but in recent decades, land water depletion, related to domestic, agricultural and industrial processes has begun to dominate. Changes in terrestrial reservoirs may also be related to climate variability: In particular, the El Nino Southern Oscillation (ENSO) has a strong impact on precipitation distribution and temporary storage of water on continents (Boening et al., 2012; Cazenave et al., 2012; Fasullo et al., 2013).

4.2.1.6 Geodynamic Processes

Land ice, glaciers and thermal expansion dominate GMSL change on decadal and longer time scales, but many other processes are relevant for local sea level changes, particularly over shorter periods. The regional patterns in sea level change are modified from the global average by changes in ocean currents, salinity, and trends in atmospheric pressure. Water mass changes between land and ocean generate patterns of sea level change, so-called fingerprints (Mitrovica et al., 2001). These redistributions of water mass lead to changes in the Earth's gravity field, lithospheric flexure and rotation. Far away from the mass loading, sea level changes can be as much as 25% greater than the global average. Proximal to retreating ice sheets, sea level drops, despite the globally averaged rise in sea level. These gravitational and rotational influences of ice loss are instantaneous, but in addition a time dependent pattern arises due to on-going visco-elastic deformation of the Earth around the location of mass change. This is observed in regions previously covered by ice during the Last Glacial Maximum (LGM), including much of Scandinavia and parts of North America (Lambeck et al., 1998; Peltier, 2004).

At the same time, more recent adjustments to recent loading changes of ice and water can have an important impact on local relative sea level. Part of this water loading change is caused by non-climate driven anthropogenic processes. For example, groundwater, oil, and natural gas extraction not only contributes to sea level change, but locally also causes large subsidence rates associated with sediment compaction, which in specific regions produce changes in relative sea level much larger than climate driven changes (Erkens and Sutanudjaja, 2015; Xu et al., 2015).

4.2.1.7 Sea Level Extremes

Superposed on gradual changes in RSL, tides, storm surges, waves and wave run-up (being the sum of wave set-up and wave swash) and other high-frequency processes (Figure 4.2) can be important locally. Understanding the localized impact of such processes requires detailed knowledge of bathymetry, erosion and sedimentation, but also a good description of the temporal variability of wind fields generating storm surges. The potential for compounding effects, like storm surge and high SLR, are of particular concern as they can contribute significantly to flooding risks and extreme events (Little et al., 2015a). These processes can be captured by hydrodynamical models (see Section 4.2.3.5).

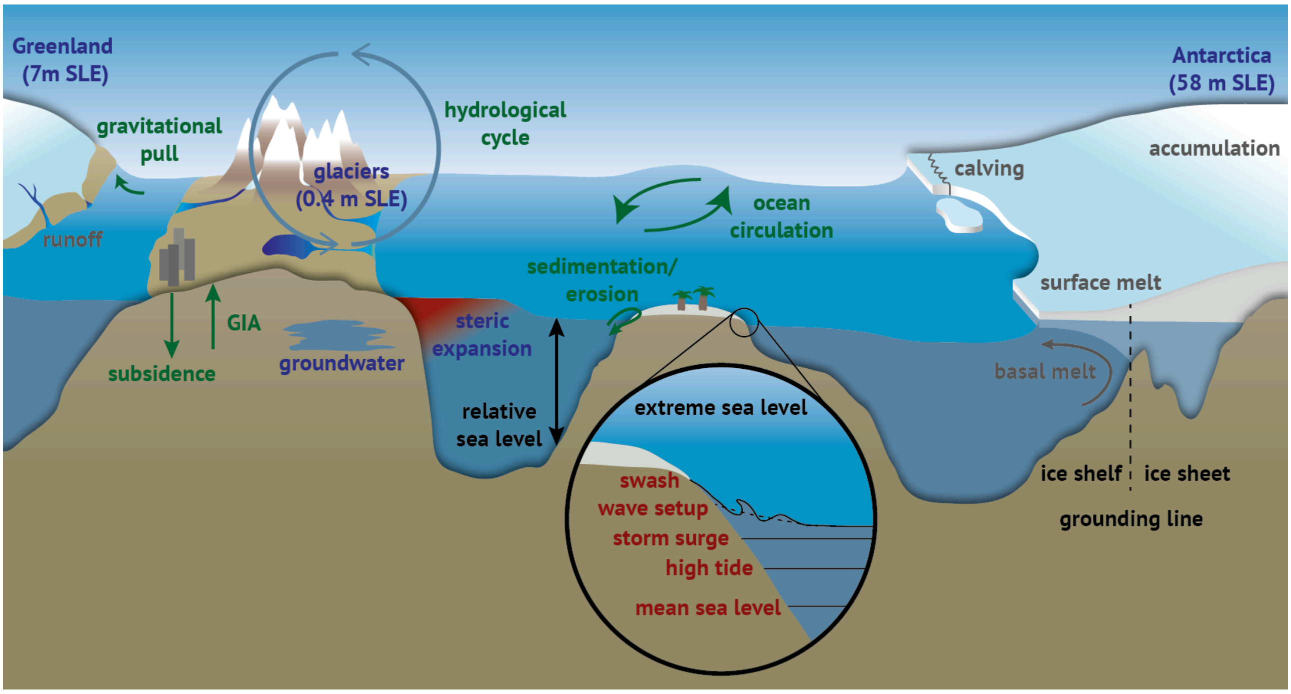


Figure 4.2: A schematic illustration of the climate and non-climate driven processes that can influence global, regional (green colours), relative and extreme sea level (red colors) along coasts. Major ice processes are shown in grey and general terms in black.

4.2.2 Observed Changes in Sea Level (Past and Present)

Past changes in sea level are important as they provide information on the size of the major ice sheets in climates different from today. Past climate intervals warmer than today, are of particular interest, because they can be used to test and calibrate process-based ice sheet models. These include the Mid Pliocene Warm Period (MPWP) around 3 Myrs BP, when according to multiple proxies of ocean temperature (Dowsett et al., 2009) global mean temperature was at times 2°C–4°C warmer than today. A second period of interest is the Last Interglacial (LIG) or Eemian around 130–115 Kyr BP, when global mean temperatures are now estimated to be 0.5°C–1.0°C higher than during the pre-industrial period (Otto-Bliesner et al., 2013; Hoffman et al., 2017) but with considerable spatial and temporal variability throughout the interval (Bakker et al., 2014; Capron et al., 2014). This is lower than the estimate of 1°C–2°C as presented in AR5 by V. Masson-Delmotte (2013). During those warm periods, sea level reconstructions indicate sea levels substantially higher than present-day, although considerably uncertainty remains.

In Section 4.2.2.1, we summarize recent advance in reconstructing these time periods in terms of climate, sea level maxima, and rates of sea level rise, and implications for future evolution of the ice sheets. In addition to periods with elevated sea level relative to modern, we consider the last deglaciation as a period of substantial and rapid ice loss. In Section 4.2.3 we discuss more recent observation of sea level changes.

4.2.2.1 Paleo Sea Level

4.2.2.1.1 Mid-Pliocene/Mid-Piacenzian warm period

The Mid-Pliocene Warm Period (MPWP) is far beyond the temporal limit of ice cores, but several geochemical techniques have been used to reconstruct Pliocene carbon dioxide concentrations from sediment archives and fossil leaves (O'Brien et al., 2014; Martínez-Botí et al., 2015), with recent estimates ranging from 300 to 350 ppmv, except the stomata based estimate by Hu et al. (2015), which finds evidence for values below 300 ppmv. Despite these relatively modest CO_2 concentrations, temperature peaked between 1.9°C to 3.6°C above pre-industrial (Haywood et al., 2016), implying high climate sensitivity (Pagani et al., 2010). Most sea level estimates for this period are considerably higher than at present. A recent compilation by Dutton et al. (2015) argues that GMSL was at least 6 m higher, but with little constraints on the maximum. The IPCC AR5 (V. Masson-Delmotte, 2013) assessed the maximum to be 14 m, with high confidence that it did not exceed 20 m. Correcting Pliocene shoreline observations for GIA (Raymo, 2011) and new insights regarding the role of dynamic topography, the vertical movement of the Earth's surface in response to mantle dynamics, Rovere et al. (2014) questioned interpretations of the sea level high stand, highlighting the ongoing uncertainty.

During the MPWP, obliquity-paced variations of up to 30 m have been reconstructed based on marine oxygen isotope data, but those are probably related to obliquity changes and not important for assessing the current changes on a shorter time scale (Naish and Wilson, 2009). Updated oxygen isotope mass balance calculations, comparing the isotopic composition of the modern and Pliocene ocean (Winnick and Caves, 2015), suggest Pliocene sea level was only ~9–13.5 m above modern, with a relatively small ~2–4.5 m contribution from the East Antarctic Ice Sheet (EAIS) in addition to the WAIS and Greenland, but the isotope approach relies on the average of multiple isotope records (Lisiecki and Raymo, 2005) with limited (~3 kyr) temporal resolution that might not capture the full range of Pliocene sea level variability. Subsequent work, using isotope enabled climate and ice sheet models to constrain the isotope mass balance problem concluded a Mid Pliocene Antarctic ice mass loss equivalent to as much as 13 m is consistent with isotope records (Gasson et al., 2016). This higher estimate implies that almost 10 m of sea level rise was contributed by East Antarctica, in line with interpretations of marine sediment cores from the East-Antarctic margin indicating substantial ice sheet variability in deep East Antarctic basins (Cook et al., 2013). Higher than present-day sea levels are also supported by sediment data (De Schepper et al., 2014) suggesting at least ice-free conditions in the Northern Hemisphere and Patagonia. A global ice-sheet modeling study by de Boer et al. (2017) suggest that the ice sheets in Greenland and Antarctica responded out of phase as a consequence of precessional orbital forcing, with a total maximum contribution of 13.3 m. The potential for interhemispheric antiphasing of ice volume (Raymo, 2006; de Boer et al., 2017) is an important emerging issue, but little work has been done to date. For example, the expansion of ice in Greenland during a MPWP high stand, would require consequently a larger contribution from Antarctica than the global average rise in GMSL. Recent Antarctic ice sheet modelling studies of maximum, mid-Pliocene ice loss (Austermann et al., 2015; Pollard et al., 2015; Yamane, 2015; DeConto and Pollard, 2016) range widely, between 5.4 and 17.8 m. An intercomparison experiment (de Boer et al., 2017) indicates that the largest uncertainty in modeling the MPWP is related to the mass balance forcing of the Antarctic ice sheet models, although the addition of new and uncertain physical processes in one model, including the influence of surface meltwater on crevassing and marine-terminating ice cliff failure (MICI) is largely responsible for the higher model estimates in Pollard et al. (2015) and DeConto and Pollard (2016).

Despite the large spread in evidence, sea level is estimated to be higher than during the LIG high stand of 6–9 m above present-day (*medium confidence*), but the mechanisms leading to the high sea level are only poorly understood.

4.2.2.1.2 Last interglacial

Dutton et al. (2015) present a revised review of Eemian sea level based on geological indicators, suggesting that global mean sea level was 6 to 9.3 m higher than today. This is in line with an earlier estimate by Kopp et al. (2009) , but slightly higher than AR5's central estimate of 6 m. Rohling (2017) consider the possibility that the Penultimate Glaciation leading up to the Last Interglacial (LIG) had a very different distribution of land ice than the Last Glacial Maximum (LGM), as commonly assumed when GIA corrections are made on LIG sea level estimates, possibly having consequences for estimates of the high stand(s). Consistent with AR5, we conclude that there is *high confidence* that LIG sea level did not exceed 10 m.

Furthermore, due to ongoing uncertainties in individual LIG shoreline data and associated GIA corrections, relative contributions of the Greenland and Antarctic Ice Sheets to LIG sea level remain uncertain, as does the exact timing of peak sea level within the interglacial (Dutton et al., 2015). In particular, there is controversy on the shape of the sea level curve throughout the Eemian (Rovere et al., 2016). Early studies argue for a double peak in the GMSL (Kopp et al., 2013), whereas more recent work (Dutton and Lambeck, 2012) suggest that part of the double peak shape may be caused by uncertainties in the GIA corrections. The relevance of solving this dispute for the present-day evolution of the ice sheet is that it may help us to understand the physical mechanisms of on-going changes in West-Antarctica and the relative contributions to LIG sea level rise from Antarctica and Greenland.

Simulations with a coupled regional climate model and ice sheet model for Greenland indicate a Greenland contribution of only up to 50 mm SLC per century (Helsen et al., 2013) and a total contribution to LIG sea level of as little at 0.75 m (Quiquet et al., 2013) and not likely more than 2.5 m (Helsen et al., 2013; Stone et al., 2013; Colleoni et al., 2014). Moreover modelling studies indicate a late peak for the Greenland contribution around 123–122 ka BP (e.g., Goelzer et al., 2013; Helsen et al., 2013; Quiquet et al., 2013). Besides ice sheet modelling studies, ice core analyses and internal ice layer imaging by radar (Dahl-Jensen et al., 2013) indicate limited ice loss for the Greenland ice sheet. Ice core results are however not fully compatible with a limited retreat, given the large increase in temperature (Dahl-Jensen et al., 2013; Landais, 2016; Yau, 2016). This suggests either a rather insensitive Greenland ice sheet to temperature changes or an overestimation of the temperature from the oxygen isotope records. The contribution of thermal expansion to LIG GMSL rise was modest (0.35–0.4 m; McKay, 2011; Goelzer, 2016) suggesting that significant parts of the Antarctic Ice Sheet contributed to the Eemian high-stand, particularly early in the interglacial before 125 ka. At the same time new geological evidence (Bierman et al., 2014) used cosmogenic 10 Be and 26 Al of marine sediments to argue that large ice caps existed in east Greenland during the last 7.5 Myr. Data from 10 Be and 26 Al measurements of sediments below the ice suggest extensive ice-free conditions in Greenland's interior (Schaefer et al., 2016), but the duration and frequency of such events are poorly known. Whether these geological findings are compatible strongly depends on the shape of the LIG ice sheet, which remains poorly constrained due to uncertainties in mass balance. Spatial patterns of retreat vary strongly among the existing model studies using different mass balance forcings (Colleoni et al., 2014).

In summary, we have *high confidence* that LIG sea level was 6–9 meters higher than today, but ongoing uncertainties in the observational evidence and ice sheet models continue to hamper conclusions regarding the rates of sea level rise, or the relative contributions from the loss of Greenland versus Antarctic ice during the warm interval.

4.2.2.1.3 Last deglaciation

Sea level rise during the Last Deglaciation (~19–11 ka) was mainly driven by the retreat of Laurentide and Fenno-Scandanavian ice sheets that no longer exist. However, there is substantially more evidence available for this period than for the LIG or MPWP, and it is the last period in the geological past when ice sheets melted rapidly. Therefore, data from this period may reveal information on the physical processes causing the ice sheet retreat, which may be difficult to retrieve from present-day observations. For example, recent evidence of keel-plough marks on the sea-floor support evidence for marine ice-cliff instability (see Section 4.2.3.1) as a process playing a role in the deglaciation of Pine Island Bay, Antarctica (Wise et al., 2017), and detailed information on the timing and pace of Antarctic Ice Sheet retreat in the Ross Sea have been used to quantitatively judge ice sheet model performance (e.g., Pollard et al., 2015). However, it is important to note that the retreat of ice sheets during the Holocene occurred during climatic conditions generally colder than today, so the mechanisms of retreat maybe be different from those that will dominate the future.

4.2.2.2 Global Mean Sea Level Changes During the Instrumental Period

Sea level observations on more recent timescales have mostly relied on tide gauge measurements. This record, extending to around 1700 in some locations, provides insight into historic sea level trends, but since the early 1990s, the emergence of satellite observations has advanced our knowledge considerably through a combination of near global ocean coverage and high spatial resolution and more detailed monitoring of ice mass changes. Since 2002, high precision gravity measurements provided by the GRACE mission (Gravity Recovery and Climate Experiment) show the loss of land ice in Greenland and Antarctica (e.g., Velicogna and Wahr, 2005; Velicogna and Wahr, 2006) confirming independent assessments of ice sheet mass changes based on satellite altimetry and InSAR measurements, combined with regional climate models to calculate SMB (Thomas et al., 2006; Rignot et al., 2008). As a result, an improved understanding of the magnitude and relative contributions of the different processes causing sea level change has emerged since AR5, particularly the increasing contribution of the ice sheets.

4.2.2.2.1 Tide gauge records

The number of tide gauges has increased over time from only a few, in northern Europe in the 18th century, to more than 2000 today along the world coastlines of continents and islands. Because of their location and limited number, tide gauges sample the ocean sparsely and non-uniformly, with a bias towards continental coastlines (only a small number of them are located on islands) and the Northern Hemisphere. Tide gauges are grounded on land and are affected by the vertical motion of Earth's crust caused by both natural processes (e.g., GIA, tectonics and sediment compaction; Tide gauges are grounded on land and are affected by the vertical motion of Earth's crust caused by both natural processes (e.g., GIA, tectonics and sediment compaction; Wöppelmann and Marcos, 2016) and anthropogenic activities (e.g., groundwater depletion, dam building or settling of landfill in urban areas; (e.g., Raucoules et al., 2010). The sea level records can be corrected for this vertical land motion (VLM) by collocated stations of the Global Positioning System (GPS) network (Santamaria-Gomez et al., 2016). Church et al. (2013) summarized the different strategies developed to account for both the inhomogeneous space and time coverage of tide gauge data and the corrections by VLM. On this basis, they estimated the sea level trend and acceleration over the period 1900–2010. They concluded that it is *very likely* that the long-term trend in GMSL from tide gauge records is 1.7 (1.5 to 1.9) mm yr^{-1} between 1900 and 2010 with a *likely* average acceleration over the 20th century between –0.002 to 0.019 mm yr^{-2}.

Since AR5 two new approaches have been developed to estimate the GMSL. The first one uses a new statistical approach with a Kalman smoother which combines tide gauge records with spatial fingerprints of ocean dynamics, GIA and ice melting to account for the inhomogeneous distribution of tide gauges (Hay et al., 2015). The second approach uses ad hoc corrections to tide gauge records with an additional fingerprint from the changes in terrestrial water storage to account for the inhomogeneous distribution in tide gauges (Dangendorf et al., 2017) and it accounts for VLM using both GPS measurement and a new method based on satellite altimetry (Wöppelmann and Marcos, 2016; Santamaría-Gómez et al., 2017). Both methods lead to GMSL increase rates that are significantly lower than AR5 estimates before 1990. Their long-term trend since 1900 is also smaller than AR5 estimates by 0.4 mm yr^{-1} (see Figure 4.3). Different arguments including biases in the tide gauge datasets (Hamlington and Thompson, 2015), biases in the averaging technique and biases in the VLM correction (Dangendorf et al., 2017) have been proposed to explain these differences with earlier AR5 estimates. These arguments for the difference do not rule out the more recent GMSL estimates or previous AR5 estimates. They rather show that the uncertainty in GMSL reconstructions is larger than previously thought and is still poorly understood from a tide gauge observational perspective. Hence, on the basis of this we conclude that it is *very likely* that the long-term trend in GMSL estimated from tide gauge records is 1.5 (1.1–1.9) mm yr^{-1} between 1900 and 2012 for a total sea level rise of 0.19 (0.17–0.21) m. In addition, we conclude with *high confidence* that sea level has accelerated over the 20th century as four of five reconstructions extending back to at least 1900 show an acceleration (Jevrejeva et al., 2008; Church and White, 2011; Ray and Douglas, 2011; Hay et al., 2015; Dangendorf et al., 2017). The estimates of the acceleration ranges between 0.002–0.019 mm yr^{-1}. The range is large and could be improved (Watson, 2016).

4.2.2.2.2 Satellite altimetry

High precision satellite altimetry started in October 1992 with the launch of the TOPEX/Poseidon and Jason series of spacecraft. Since then, 11 satellite altimeters have been launched providing nearly global sea level measurement (up to ±82° latitude) at different temporal sampling (from 3 to 35 days) over more than 25 years. Unlike tide gauges, altimetry measures sea level relative to a geodetic reference frame and thus is not affected by VLM. But altimetry measurement can be affected by instrumental biases, in particular in the early altimetry era when TOPEX/Poseidon was flying alone. Since AR5, several studies using two independent approaches based on tide gauge records (Watson et al., 2015), the sea level budget closure (Chen et al., 2017; Dieng et al., 2017) identified a drift of 1.5 (0.4–3.4) mm yr^{-1} in TOPEX A over the period January 1993 to December 1998. Accounting for this drift leads to a revised rate of the global MSL from satellite altimetry of 3.0 mm yr^{-1} (2.4–3.6) over the period 1993–2015 instead of 3.3 mm yr^{-1} (2.7–3.9) as stated in the AR5. Hence, a revised estimate of the satellite altimetry GMSL record now shows an acceleration of 0.084 (0.059–0.090) mm yr^{-1} over 1993–2015 (Watson et al., 2015; Nerem, 2018). This acceleration is mostly due to an increase in Greenland mass loss since the 2000s (Chen et al., 2017; Dieng et al., 2017) and a slight increase in all other components probably partly due to the recovery from the Pinatubo volcanic eruption in 1991 (Fasullo et al., 2016) and partly due to the increased GHG concentrations (e.g., Slangen et al., 2016).

4.2.2.3 Contributions to GMSL Change During the Instrumental Period

In this section, we estimate the observed contributions to the GMSL rise and assess the closure of the sea level budget. In addition, to assess our understanding of the causes of observed changes and our confidence in projecting future changes, we compare observational estimates of contributions with results derived from AOGCM experiments, beginning in the mid-19th century, forced with estimated past time-dependent anthropogenic changes in atmospheric composition, and natural forcings due to volcanic aerosols and variations in solar irradiance. The period since the mid-19th century and these simulations are often referred to as ‘historical’.

4.2.2.3.1 Thermal expansion contribution

Thermal expansion is a major contribution to the rate of global mean sea level rise (GMSLR), about 0.5 to 1.1 mm yr^{-1} for 1971–2010 and 0.8 to 1.4 mm yr^{-1} for 1993–2010 (Church et al., 2013; Rhein et al., 2013), *very likely* due to anthropogenic warming of the ocean (Bindoff et al., 2013; Church et al., 2013; Rhein et al., 2013; Slangen et al., 2014c; Gleckler et al., 2016).

Thermal expansion estimates can be directly calculated from in situ ocean observations and also through ocean syntheses that rely on assimilation of data into numerical models. Full-depth, high-quality and unbiased ocean temperature profiles with adequate metadata and spatio-temporal coverage are required to estimate thermal expansion and to understand its causes, however, the global observing system is not ideal (Abraham et al., 2013; Good, 2017). Several factors can introduce uncertainty (Palmer et al., 2010), the largest being the choice of mapping methods for estimates in the upper 700 m during 1970 (Boyer et al., 2016).

For those observations the thermal expansion contribution ranges from 24% to 42% of the GMSLR rate for 1993–2015, depending on XBT correction (Pittman et al., in prep.). If we consider bias-corrected contributions including all recommended factors (National Centers for Environmental Information, 2017), results lie in the upper-range (~40%). Further coordinated evaluation/refinement of mapping methodologies (e.g., [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added]) and XBT bias corrections (e.g., Cheng et al., 2016a) will help to reduce uncertainty.

Greater agreement among estimates is found for more recent data from 2006–2007, when the array of Argo profiling floats reached its targeted near-global (60ºN to 60ºS) coverage in the upper 2,000 m (Roemmich et al., 2015; Riser et al., 2016; Schuckmann et al., 2016; Wijffels et al., 2016). During 2006–2015, global ocean heat gain of 0.50–0.65 W m^{-2} (with an additional 0.15 W m^{-2} due to deep ocean warming and ocean areas not sampled by Argo), equally divided between 0–500 m and 500–2000 m, with a broad maximum between 700–1400 m. Most of this decadal heat gain (75% to 98%) occurred in the Southern Hemisphere, with a zonally-averaged maximum at 40ºS (Roemmich et al., 2015; Wijffels et al., 2016), largely due to thermal expansion associated with a volumetric increase of subtropical mode waters (Desbruyères et al., 2017).

Since AR5, evaluation for global thermal expansion rates (GThSLR) are 0.76 mm yr^{-1} for 1970–2015, 1.20 ± 0.23 mm yr^{-1} for 1992–2014 (Chambers et al., 2017) and 1.38 ± 0.16 mm yr^{-1} for 2002–2014 (Rietbroek et al., 2016).

Historical GMSL rise due to thermal expansion simulated by CMIP5 models is shown in Table 4.1. For models that omit the volcanic forcing in their control experiment, the imposition of the historical volcanic forcing on the historical climate results in a spurious time mean negative forcing and a spurious persistent ocean cooling related to the control climate (Gregory, 2010; Gregory et al., 2013a). The magnitude of this effect is estimated from historical natural-only simulations and then used to correct the historical simulations (Slangen et al., 2016; Slangen et al., 2017c). This approach is a refinement of the methodology used in AR5 where a constant correction of 0.1 mm yr^{-1} was applied to all the model results. The model spread in thermal expansion is larger than the observational uncertainties (Cheng et al., 2016b; Gleckler et al., 2016). This spread is due to uncertainty in radiative forcing and uncertainty in the modelled climate sensitivity and ocean heat uptake efficiency (Melet and Meyssignac, 2015). The ensemble mean of modelled thermal expansion provides a good fit to the observations within the uncertainty ranges of both models and observations (Roemmich et al., 2015; Riser et al., 2016; Schuckmann et al., 2016; Wijffels et al., 2016; Slangen et al., 2017b). The improved observed and modelled estimates of thermal expansion, the good agreement between both estimates, and the improved understanding of the spread between modelled estimates give *high confidence* in the simulated thermal expansion using climate models. It also provides *high confidence* in the ability of climate models to project future thermal expansion.

4.2.2.3.2 Ocean mass observations from GRACE

Since 2002, it is possible to directly estimate the ocean mass changes with space gravimetry data from the Gravity Recovery and Climate Experiment (GRACE) mission. The ocean mass changes correspond to the sum of land ice and terrestrial water storage changes so the direct measurement from GRACE provides an independent estimate of these contributions. Since AR5, owing to the extended record of GRACE gravity measurements (over 15 years), improved understanding of GRACE gravity data and methods for addressing GRACE limitations (e.g., noise filtering, leakage and low-degree spherical harmonics), and improved knowledge of geophysical corrections (e.g., GIA), GRACE-derived ocean mass rates show increased consistency (Table 4.1). Recent estimates (Dieng et al., 2015b; Reager et al., 2016; Rietbroek et al., 2016; Chambers et al., 2017) report a global ocean mass increase of 1.7 (1.4 to 2.0) mm yr^{-1} over 2003–2015. The associated uncertainty arises essentially from differences in the inversion method to compute the ocean mass (Chen et al., 2013; Jensen et al., 2013; Johnson and Chambers, 2013; Rietbroek et al., 2016) and uncertainty in the GIA correction. The consistency between estimates of the global mean ocean mass on a monthly time scale has also increased since AR5 the biggest differences between monthly estimates being now of the order of 5 mm.

4.2.2.3.3 Glaciers

To assess the mass contribution of glaciers to sea level change, global estimates are required. Recent updates and temporal extensions of estimates obtained by different methods continue to provide *very high confidence* in continuing glacier mass loss on the global scale and show increased agreement on rates of mass loss during the 20th century, compared to earlier estimates reported by Vaughan (2013). Rates of early 21st century glacier mass loss on the global scale were found to be unprecedented during the observed period (Zemp, 2015).

Updates of three long-term time series were presented jointly in Marzeion et al. (2015): First, an update of the compilation of Cogley (2009) which combines geodetic and direct measurements of glacier mass change resulting in slightly lowered mass loss rate estimates, particularly within the first decade of the 21st century. This revision was mostly based on an increased number of regional-scale geodetic mass balance observations and lead to an improved agreement with the rates previously reported by Gardner (2013) obtained by orbital altimetry and gravimetry. Second, the method of Leclercq (2011), which is based on glacier length records, was revised by including the extended glacier length database presented in Leclercq (2014), and recalibrated using the mass change update of Cogley (2009) mentioned above. The increased number of glacier length records, mostly in the Arctic, resulted in a better representation of the global distribution of glaciers. In combination with the re-calibration, the mass loss rate estimates for the 20th century were considerably increased with respect to the estimate presented by Church et al. (2013). Third, based on forcing a glacier model with gridded climate observations, the estimate of Marzeion et al. (2012) was revised by updating the glacier inventory used for initialization (Pfeffer, 2014) from version 1.0 to version 4.0, and the gridded climate observations used as forcing (Harris et al., 2014) from version 3.0 to version 3.22. Both updates lead to reduced mass loss rate estimates in the 20th and early 21st century. All these three long-term time series are now agreeing with each other, within their respective uncertainties, for their entire common periods, on the global scale. The mean rates of glacier mass loss estimated during 1901 to 2010 are 0.62 ± 0.05 mm SLE yr^{-1} (update from Marzeion et al., 2012) and 0.78 ± 0.19 mm SLE yr^{-1} (update from Leclercq, 2011). During 1961 to 2010 they are 0.49 ± 0.05 mm SLE yr^{-1} (update from Marzeion et al., 2012), 0.54 ± 0.05 mm SLE yr^{-1} (update from Cogley, 2009), and 0.58 ± 0.15 mm SLE yr^{-1} (update from Leclercq, 2011). During 2003 to 2009 they are 0.78 ± 0.15 mm SLE yr^{-1} (update from Marzeion et al., 2012), 0.75 ± 0.07 mm SLE yr^{-1} (update from Cogley (2009), 0.84 ± 0.64 mm SLE yr^{-1} (update from Leclercq, 2011), and 0.70 ± 0.07 mm SLE yr^{-1} (Gardner, 2013). Disagreements between the different methods exceeding the respective uncertainty estimates remain on the regional scale.

Gravimetric estimates of glacier mass loss are available since 2002. Their uncertainty is mostly related to the difficulty of attributing observed mass changes to glaciers and other sources of mass change (such as regional land hydrology and Solid Earth signals) which is exacerbated by their small size relative to the resolution of the measurement and by their spatially heterogeneous distribution. The strong temporal variability of glacier mass change rates in combination with the still relatively short gravimetric time series and diverse periods of assessment complicates the comparison of the different estimates (e.g., Chen et al., 2013; Schrama, 2014; Dieng et al., 2015a; Reager et al., 2016; Rietbroek et al., 2016). However, these estimates tend to result in lower glacier mass change rates than those based on direct and geodetic observations, and glacier modelling. The glacier contribution estimated from GRACE ranges from 0.3 to 0.7 mm SLE yr^{-1} for different periods within 2002–2015.

Since some of the glacier melt water runoff is within endorheic basins, and retreating glaciers may lead to the formation of new lakes, glacier mass loss is not equal to ocean mass gain. While detailed estimates of these effects are available only on the regional scale (e.g., Neckel, 2014; Kääb, 2015), there are indications that the global potential of meltwater retention is only a few percent of the runoff (Haeberli W, 2013), and probably within the uncertainty range of glacier mass loss estimates. However, increasing lake levels may affect regional glacier mass change estimates obtained by gravimetry if the mass change is misattributed (Zhang, 2013).

Climate models have a spatial resolution that is too coarse to resolve the glaciers' mass balance in the narrow glacial valleys (which are typically a few tens of km wide or less). Therefore, mass balance models of glaciers are forced off-line using climate model results, and involving some kind of downscaling (e.g., Radić et al., 2014; Huss and Hock, 2015; Slangen et al., 2017a; Marzeion et al., 2018). Compared to AR5, the new model results find a reduced glacier contribution to the 20th century sea level rise of 55 ± 13 mm. This reduced contribution is mainly the result of the updated glacier inventory and improvements in the digital elevation model (Marzeion et al., 2015).

Compared to the long time series of observed glaciers mass changes from Leclercq (2011) and Marzeion et al. (2015) the modelled contribution from climate models is smaller in the beginning of the 20th century (see Figure 4.3). This is due to a large mass loss in the glaciated regions around Greenland, corresponding to a period of warming and strong glacier melt that is present in the observation-forced model and also found in observations (Bjørk et al., 2012), but not replicated by the CMIP5 models. The reason for this regional underestimation of glacier mass loss is not clear. It can be due to internal variability (Chylek et al., 2004;

Church et al., 2013; IPCC, 2014), or to a bias that was found in the atmospheric circulation of climate models, which do not reproduce warm air flow over the south of Greenland resulting in too little melt in this region (Fettweis et al., 2013). Over the second half of the 20th century, when the Cogley (2009) observations are also available, the differences between modelled and observed glacier mass change are reduced. It suggests that the cause for the discrepancy over the first half of the century is rather internal variability.

In view of the better agreement between observational estimates of glaciers mass changes (in particular in the first half of the 20th century), and the better consistency of historical model results with observational estimates, the confidence in the use of glacier models to reconstruct sea level change has increased since AR5. But the increase in confidence is limited and the overall confidence remains medium because of the still limited number of well-observed glaciers to validate models on long time scales, because of the unexplained bias in models over the first part of the 20th century, and because of the small number of model-based global glacier reconstructions.

4.2.2.3.4 Greenland and Antarctic ice sheets

Because ice sheets are remotely located, reliable observations of their ice mass changes have only been available since the advent of space observations. Three types of methods are in use to estimate ice sheet mass balance: satellite altimetry, where mass loss is estimated by direct measurements of height changes by laser or radar altimetry, in combination with climatological/glaciological models for firn density and compaction; input–output methods, where measurements of ice flow velocities estimated from synthetic aperture radar data along the margin are combined with ice thickness data, and models for accumulation and ablation on the ice sheet are used to give a net surface mass balance; space gravimetry data yield direct estimate of the mass changes (see Section 4.2.2.3.2). Vaughan (2013) concluded that the three space-based methods give consistent results. They agree in showing that the contribution of the Greenland and Antarctic ice sheets has increased since the early 1990s, partly from increased outflow induced by warming of the immediately adjacent ocean. Since AR5, up-to-date observations confirm this statement (Cazenave et al., 2018). They indicate a Greenland and Antarctica mass loss in the period 2002–2015 amounting to 265 ± 25 Gt yr^{-1} for Greenland (including peripheral ice caps), and 95 ± 50 Gt yr^{-1} for Antarctica, corresponding respectively, to 0.72 and 0.26 mm yr^{-1} global mean sea level change. A significant acceleration in mass loss rate is found for Antarctica (McMillan et al., 2014) and Greenland (Enderlin, 2014; van den Broeke, 2016). In Greenland, where substantial interannual variability in mass balance has been common throughout the satellite record, a swing between extreme melting and accumulation events from 2012 to 2013–2014 (Tedesco et al., 2016) is consistent with large recorded mass loss followed by a temporary abatement. In Greenland, the acceleration is caused by a decrease in SMB and an increasing flow and retreat of outlet glaciers (van den Broeke, 2016). In contrast, Antarctica's recent increase in mass loss is not through surface melt, but is instead mostly related to the increasing flow and retreat of outlet glaciers in the Amundsen Sea region of West Antarctica (Mouginot et al., 2014; Rignot et al., 2014). Warming ocean temperatures resulting from changes in the ocean circulation are thinning ice shelves triggering a dynamic response of the grounded ice upstream (Paolo et al., 2015).

Modelled changes in Antarctica and Greenland SMB are obtained from regional climate models or downscaled global climate models. Since AR5, global climate models are downscaled with new regional statistical techniques which account for the non-uniform distribution of SMB changes over Greenland and Antarctica (Noël et al., 2015; Favier et al., 2017; Meyssignac et al., 2017). There are no direct observational time series of Greenland and Antarctica SMB over the 20th century, but observational estimate can be obtained using atmospheric reanalysis data to force regional climate models. The contribution of Greenland and Antarctica SMB changes to GMSL estimated from global climate model amounts is given in Table 4.1. For Greenland, the climate model based estimates agree with reanalyses-based estimates and direct observations in showing an abrupt increase in SMB contribution to GMSL since 1990 (Van Angelen et al., 2014). Before 1940 the reanalysis-based Greenland SMB estimates show a significantly larger contribution to GMSL than the climate model based estimates. As for glaciers, it is attributed to an increase in air temperatures in and around Greenland over the period 1900–1940, which led to increased melt in Greenland (Bjørk et al., 2012; Fettweis et al., 2017) and surrounding glaciers. This difference can be due to internal climate variability that is not supposed to be captured by climate models, or a bias in atmospheric circulation in climate models (Fettweis et al., 2013), or an issue with the spatial pattern of the historical aerosol forcing. For Antarctica, reanalyses-based estimates only provide estimates since 1979 because atmospheric reanalyses are not reliable over Antarctica before (Favier et al., 2017). Over 1979–2015 climate model based estimates of Antarctica SMB agree with reanalyses-based estimates showing a small contribution to GMSL rise. For the more recent period 2005–2015 the Antarctic contribution to sea level as measured by GRACE and Input Output Method is estimated to be 0.42 ± 0.1 mm yr^{-1}, whereas individual years driven by surface mass balance processes vary between -0.4 and +0.7 mm yr^{-1} (Cazenave et al., 2018). The largest uncertainty in trend estimates by GRACE is caused by the uncertainty in the GIA correction. Most studies indicate a mass loss from the Antarctic Peninsula and West Antarctica, for East Antarctica the signal is small compared to the uncertainties. These recent estimates confirm that the mass loss of Antarctica accelerated over the last 10 years with respect to earlier periods (see Table 4.1).

4.2.2.3.5 Contributions from water storage on land

Large-scale natural changes in land water storage, defined as snow, surface water, soil moisture, and groundwater storage, excluding glaciers, contribute to observed changes in sea level on annual to centennial timescales (Döll et al., 2016; Reager et al., 2016; Wada et al., 2017). Direct anthropogenic intervention on land hydrology also contributes to sea level changes at these time scales. It includes human transformations of Earth's surface which impacts continental patterns of river flow and water exchange between land, atmosphere, and ocean, ultimately affecting global sea level variations (Döll et al., 2016; Wada et al., 2016). It includes also the massive impoundment of water in reservoirs and artificial lakes which reduces the outflow of water to the sea, while river runoff increases due to increased groundwater mining, wetland and endorheic lake storage losses, and deforestation. Overall, the combined effects of direct anthropogenic processes have reduced land water storage, increasing the rate of sea level rise (SLR) by 0.15–0.24 mm yr^{-1} during the last decade (Wada et al., 2016; Wada et al., 2017; Cazenave et al., 2018; Scanlon et al., 2018). The AR5 considered the effect of anthropogenic changes in terrestrial water storage (primarily filling of reservoirs and groundwater mining) on sea level but natural fluctuations were excluded due to poor knowledge of their change and the assumption that such changes would be small on decadal timescale. Recently, Dieng et al. (2015a), Dieng et al. (2017), and Reager et al. (2016) showed that climate-driven changes in water storage (e.g., soil moisture and groundwater) have a large impact on global sea level variations over decadal timescales. Two approaches based on hydrological models and GRACE observations enable to estimate the net land water storage contribution. They provide different estimates of associated sea level rate that range -0.33 to 0.23 mm yr^{-1} during the period of 2002–2014/15. According to GRACE, the net TWS change (i.e., not including glaciers) over the period 2002–2014 shows a negative contribution to sea level of -0.33 mm yr^{-1} and -0.21 mm yr^{-1}, as shown by Reager et al. (2016) and Scanlon et al. (2018) respectively while hydrological models estimate slightly positive trends over the same period. Because of these significantly different estimates, we have *medium confidence* in the TWS contribution to the current sea level rise. Although the time period is not the same as in AR5, this estimate show a discrepancy with the AR5 estimate of 0.38 mm yr^{-1} over 1993–2010. This discrepancy is essentially due to the consideration of natural changes in land water. Natural land water storage *likely* has large decadal variability (Reager et al., 2016; Dieng et al., 2017) and its contribution to sea level will probably change in the coming decades.

4.2.2.3.6 Budget of GMSL change

Drawing on previous sections, the budget of GMSL rise (Table 4.1, Figure 4.3) is assessed over different periods. As in Church et al. (2013), we assess the budget with models and observations. We consider 4 periods: 1901–1990 (the 20th century, excluding the period after 1990 when ice-sheet contributions to GMSL rise have increased), 1971–2015 (when ocean observations are sufficiently accurate to estimate the global ocean thermal expansion and when systematic glacier reconstructions start), since 1993–2015 (when precise satellite altimetry begin) and 2005–2015 (when Argo and GRACE data are both available). The period 2005–2015 is only 10 years long and can be affected by internal climate variability, which is not externally forced and is therefore not expected to be reproduced in AOGCM historical experiments. This can explain part of the discrepancy between the observed and the modelled GMSL rise budget over this period. For the contribution from land water storage, we use the estimated effect of human intervention, neglecting climate-related variation until 2005. From 2005 on, we use the total land water storage estimated with GRACE. Before 2005, climate related variations in land water are negligible because their amplitude is very small on multi decadal timescales.

For 1993–2015 and 2005–2015, allowing for uncertainties, the observed GMSL rise is consistent with the sum of the observationally estimated contributions. Over the period 1993–2015 the two largest terms are ocean thermal expansion (accounting for 42% of the observed GMSL rise) and glacier mass loss (accounting for a further 22%). Compared to AR5 the extended observations allow us now to identify an acceleration in the observed sea level rise over 1993–2015 and to attribute this acceleration mainly to Greenland ice loss with also a small acceleration in Antarctica ice loss (Velicogna et al., 2014; Harig and Simons, 2015; Chen et al., 2017; Dieng et al., 2017). Since 2005, land ice, collectively from glaciers, and the ice sheets, is now becoming the most important contributor to GMSL rise over the thermal expansion with mountain glaciers and ice caps contributing 21% and ice sheets 34% (A. Cazenave, 2018). Over the periods 1993–2015 and 2005–2015 sea level components are also consistent within uncertainties at monthly-scales with the total observed sea level with significantly smaller uncertainties during the period 2005–2015 when Argo data have global distribution and GRACE data are available. This agreement represents a significant advance since the AR5 in physical understanding of the causes of past GMSL change and provides an improved basis for the evaluation of models. It also gives *high confidence* that the current ocean observing system is capable of resolving the rate of sea level rise and its components.

Before 1990, observations are not sufficient to confidently estimate the ice sheet mass balance; before 1971, the space and time sampling of ocean observations are not sufficient to estimate the global ocean thermal expansion. For these reasons, it is difficult to assess the closure of the GMSL rise budget over 1900–1990 and 1971–2015.

For the period 1971–2015 the thermal expansion of the ocean represents 40% of the observed GMSL rise while the glaciers contribution represents 30%. This is a slightly smaller contribution from glaciers than indicated in the AR5. If we add the Greenland ice sheet contribution and the Antarctic surface mass balance then the sum of the contributors to sea level is in agreement with the low end observed sea level rise estimates over 1971–2015 (Frederikse et al., 2018). This result suggests that the contribution of Antarctica ice sheet dynamics to sea level rise has been small, if any, before the 1990s.

Since AR5, extended simulations along with recent findings in observations and improved model estimates allow for a new more robust and more consistent comparison between sea level estimates with climate models and observed sea level. Compared to AR5, the glacier contribution contains an updated glacier inventory and improvements to the digital elevation model, which have caused a decrease in the modelled estimated 20th century glacier contribution (Marzeion et al., 2015). The Greenland SMB is estimated with a new regional SMB-component downscaling technique, which accounts for the regional changes in Greenland SMB (Noël et al., 2015). In addition, recent groundwater extraction estimates (e.g., Doell et al., 2014) were used for the land water storage contribution. They tend to be lower than the values included in AR5. This is in agreement with other recent publication, showing that only 80% of the extracted groundwater ultimately reaches the ocean (Wada et al., 2016). When all the new estimated contributions are combined, there is a large gap between the observations and the models before 1990, and only 50 ± 30% of the observations (mean of five tide gauge reconstructions) can be explained by the models for the period 1901–1920 to 1970–1990 (Slangen et al., 2017c). The gap is essentially explained by a bias in modelled Greenland SMB, and glacier ice loss around Greenland in the early 20th century (see previous sections). This bias is potentially due to the internal variability of the climate system which is not expected to be in phase in climate models. When this bias is corrected, the explained percentage increases to 75 ± 38% for the mean of the five reconstructions (Slangen et al., 2017c). Compared to the individual reconstructions, the bias-corrected simulations agree best with the Hay et al. (2015) and Dangendorf et al. (2017) reconstructions, explaining 92% of the observed change.

For the more recent satellite altimetry period (from 1993–2015), the percentage explained by the simulations is 102 ± 33% (105 ± 35% when bias corrections are included), effectively closing the sea level budget for this period (Slangen et al., 2017c). In this later period, the uncertainties in the observations are smaller as the data resolution is higher, both spatially and temporally. Compared to AR5, the improved ability of climate models to reproduce the 20th century sea level changes due to thermal expansion, glacier mass loss and ice sheet surface mass balance gives *high confidence* in climate models to project future changes of these contributors to sea level. Since AR5 the ice sheet dynamics contribution has increased, but it remains relatively small up to present and the closing of the sea level budget do not test the reliability of ice-sheet models in projecting future rapid dynamical change.

Table 4.1: Budget of GMSL change.

Source	1901–1990	1971–2015	1993–2015	2005–2015
Observed contribution to GMSL	1.53 (0.96 to 2.11)	2.06 (1.76 to 2.36)	3.07 (2.70 to 3.44)	3.5 (3.3 to 3.7)
Thermal expansion	--	0.78 ± 0.28	1.3 ± 0.4	1.3 ± 0.4
Glaciers	0.77 ± 0.23	0.65 ± 0.33	0.65 ± 0.15	0.74 ± 0.1
Greenland SMB	0.11 ± 0.10	0.02 ± 0.01	0.29 ± 0.xx	0.46 ± xx
Greenland Ice sheet dynamics	--	0.19[e]	0.19 ± 0.xx	0.30 ± 0.xx
Antarctica SMB +Ice sheet dynamics	--	0.22	0.25 ± 0.1	0.42 ± 0.1
Land water storage	–0.12[c]	–0.07[c]	0.09[c]	–0.05 ± 0.28[d]
Ocean mass	--	--	1.47 ± 0.xx[b]	2.3 (2.11 to 2.49)[a]
Total contributions	--	--	2.77 ± 0.xx	3.17
Modelled contributions to GMSL rise				
Thermal expansion	0.32 (0.04 to 0.60)	0.97 (0.45 to 1.48)	1.48 (0.86 to 2.11)	1.52 (0.96 to 2.09)
Glaciers	0.53 (0.38 to 0.68)	0.73 (0.50 to 0.95)	0.99 (0.60 to 1.38)	1.10 (0.64 to 1.56)
Greenland SMB	–0.02 (–0.05 to 0.02)	0.03 (–0.01 to 0.07)	0.08 (–0.01 to 0.16)	0.12 (–0.02 to 0.26)
Antarctic SMB	–0.02 (–0.07 to 0.03)	–0.10 (–0.23 to 0.03)	–0.14 (–0.35 to 0.06)	–0.16 (–0.40 to 0.08)
Total including land water storage and ice sheet dynamics	0.69 (0.18 to 1.20)	1.78 (0.72 to 2.69)	2.99 (1.69 to 4.30)	3.38 (1.98 to 4.78)
Residual	0.94 (0.42 to 1.44)	0.27 (–0.64 to 1.33)	0.32 (–0.98 to 1.63)	0.31 (–1.08 to 1.71)

Notes:
Glaciers excluding Antarctic Peripheral glaciers.
(a) Direct estimate of ocean mass from GRACE
(b) Estimated from the sum of land ice melt and land water storage changes
(c) Only direct anthropogenic contribution
(d) Including both the direct anthropogenic contribution and the climate variability
(e) Estimated from the total estimate of Greenland ice sheet mass loss from Kjeldsen (2015) corrected for the Greenland SMB.

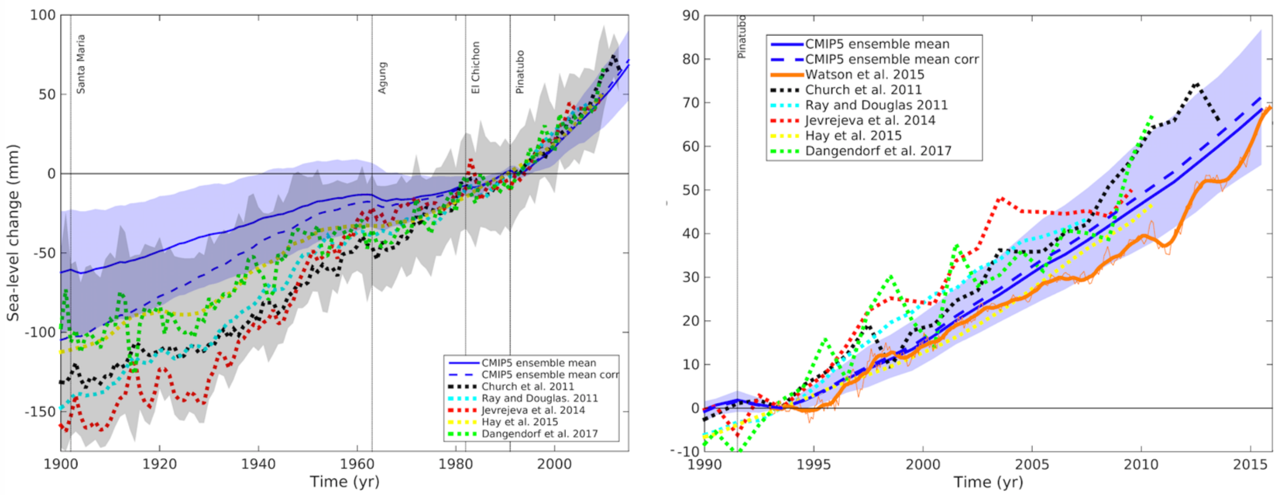


Figure 4.3: Comparison of modelled (as in Section 4.4.2.6) and observed global mean sea level change since 1900 (a) and since 1993 (b). The average estimate of 12 CMIP5 AOGCM simulations is shown in blue and estimates from observations in other colours, with the 17%–83% *likely* range shaded and calculated according to the procedures in Church et al. (2013). The average of the 12 model estimates corrected for the bias in glaciers mass loss and Greenland surface mass balance in the 1930s (see text) is shown in dashed blue. All curves in (a) and (b) are shown with zero time-mean over the period 1980–2000. Updated from Slangen et al. (2017c).

4.2.2.4 Regional Sea Level Changes During the Instrumental Period

Sea level does not rise uniformly. Observations from tide gauges and satellite altimetry (Figure 4.4) indicate that sea level shows substantial regional variability at decadal to multi-decadal time scales. These regional changes are essentially due to changing winds, air-sea heat and freshwater fluxes and the addition of melting ice into the ocean which alter the ocean circulation. Observations with ocean models, ocean reanalysis and sea level reconstructions agree in showing that the sea level patterns over the last half of the 20th century fluctuate in space and time in response to variability modes of the coupled ocean-atmosphere system such as ENSO, the NAO, and the PDO (Frankcombe et al., 2013; Nidheesh et al., 2013; Palanisamy et al., 2015a; Carson et al., 2017; Han et al., 2017; Nidheesh et al., 2017)

The variability and trends in sea surface heights (SSH) observed during the recent altimetry era or reconstructed over the past decades are largely dominated by the steric effect except in shallow shelf seas where the mass effect is of the same order of magnitude as the steric effect and at high latitudes (>60°N and <55°S) where the mass effect dominates. The steric sea level signal is essentially due to temperature changes. Salinity changes play only a local role, but this role can be sizeable in several regions in particular in the North Atlantic, in the Arctic and in the Southern Ocean. The observed steric sea level variability and trends are essentially forced by surface wind stress anomalies in particular in the tropics where the SSH variability and trends are the most intense over the last two decades. The buoyancy forcing plays also a sizeable role but of smaller amplitude and more uniformly distributed. In general, on average over the ocean, the buoyancy forcing effect on SSH trends is positive which reflects the penetration of heat into the ocean and the global warming of the ocean. While the buoyancy fluxes only are responsible for the total heat that enters the ocean and the associated global mean sea level rise, both ocean transport divergences caused by wind stress anomalies and the non-uniform buoyancy forcing (essentially at mid to high latitude) are responsible for the regional distribution of the heat within the ocean and thus for the regional sea level departures around the global mean.

Over the Pacific and Indian Ocean, sea level trend patterns since the 1970s are driven primarily by surface winds associated with the ENSO, IPO and NGPO modes in the Pacific and with the ENSO and IOD modes in the Indian ocean. Over the Atlantic, the NAO-associated sea level patterns exhibit a dipole structure in the North Atlantic basin. In the basin interior, surface heat fluxes are suggested to be the major force for the decadal sea level patterns due to AMOC variations. Along the eastern boundary, longshore winds and coastal Kelvin waves are the primary causes for the coherent sea level changes. Along the west boundary (US east coast), some studies demonstrate the importance of interior wind stress curl and local wind over the shelf in driving decadal sea level variability, whereas others argue for the importance of AMOC and Gulf Stream variations. A 20- to 30-year decadal sea level signals observed in both the North and South Atlantic are associated with AMOC variations and oceanic Rossby waves. Over the Arctic, winds and to a lesser degree inverse barometer effects are important for driving the sea level variability, and in the Norwegian Sea coastal signals propagating from the eastern boundary of the North Atlantic also contribute. Finally, in the Southern Ocean, the SAM can have a significant influence on sea level in particular in the Indian and Pacific sectors, with weak influence over the tropics compared to the PDO, ENSO and IOD. Zonal asymmetry in SAM-associated winds might have contributed to the asymmetry of decadal sea level variations in the Southern Ocean during most of the twentieth century.

As for GMSL, net regional sea level changes can be estimated from a combination of the various contributions to sea level change. The contributions from dynamic sea level, atmospheric loading, glacier mass changes and ice sheet SMB can be derived from CMIP5 climate model outputs either directly or through downscaling techniques (Perrette et al., 2013; Kopp et al., 2014; Slangen et al., 2014b; Bilbao et al., 2015; Carson et al., 2016; Meyssignac et al., 2017). The contribution from groundwater depletion, reservoir storage and dynamic ice sheet mass changes are not simulated by climate models over the 20th century and has to be estimated from observations. The sum of all contributions including the GIA contribution, provides a modelled estimate of the 20th century net regional sea level changes which can be compared with observations from satellite altimetry and tide-gauge records.

There is a general agreement between the modelled regional sea level and tide gauge records in terms of inter-annual to multi-decadal variability over 1900–2015. But, as for GMSL, climate models tend to systematically underestimate the observed sea level trends from tide gauge records, particularly in the first half of the 20th century. This underestimation is essentially explained by the bias in modelled Greenland SMB, and glacier ice loss around Greenland in the early 20th century (see previous sections). The correction of this bias lead to an improved explanation of the spatial variability in observed sea level trends by climate models. Climate models indicate that the spatial variability in sea level trends observed by tide-gauge records over the 20th century is dominated by the GIA contribution and the steric contribution over 1900–2015. Locally all contributions to sea level changes are important as any contribution can cause significant local deviations; for example, the groundwater depletion around India is responsible for the low 20th century sea level rise in the region.

These results show the ability of models to reproduce the 20th century regional sea level changes due to GIA, thermal expansion, glacier mass loss and ice sheet surface mass balance. It gives *high confidence* in climate models to project future regional changes associated with these contributors to sea level. The other contributions to 20th century sea level, including the growing ice sheet dynamics contribution, have not been simulated so far by climate models. Thus the ability of models to reproduce observed past changes has not been tested so far.

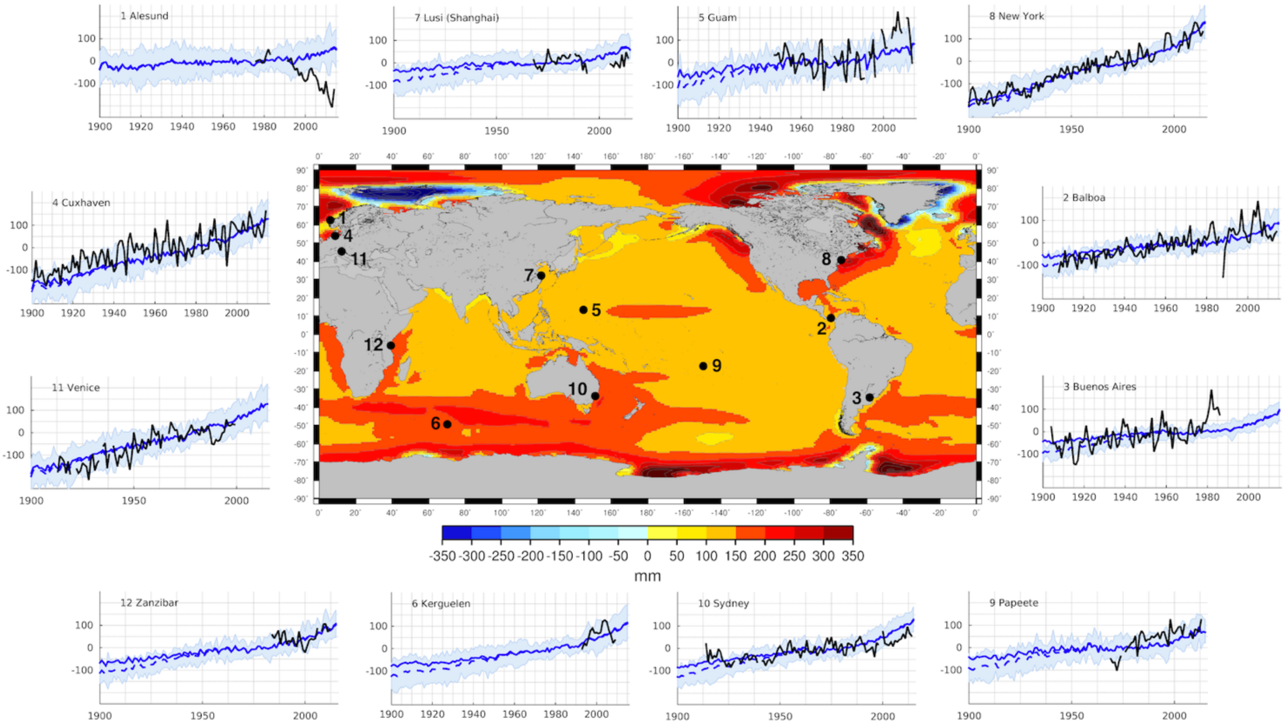


Figure 4.4: Map of rates of change in modelled relative sea level for the period 1901–1920 to 1996–2015 from AOGCMs inputs. Also shown are relative sea level changes (black lines) from selected tide gauge stations for the period 1900–2015. For comparison, the estimate of the modelled relative sea level change at the tide gauge station from AOGCMs is also shown (blue plain line for the model estimates and blue dashed line for the bias corrected model estimates) with each tide gauge time series. The relatively large, short-term oscillations in local sea level (black lines) are due to the natural internal climate variability. Tide gauge records have been corrected for vertical land motion not associated with GIA when estimates of this motion were available in the literature, for instance, New York, Balboa and Shanghai (Meyssignac et al., 2017). [PLACEHOLDER FOR SECOND ORDER DRAFT: numbering to be aligned with later figures]

4.2.2.5 Local Coastal Sea Level

Since local coastal sea level is affected by global, regional and coastal scale features and processes, it may differ substantially from the global average. At the coast, sea level change is additionally affected by wave run up, tidal level, sea level pressure (SLP), the dominant modes of climate variability (Section 4.2.2.5), seasonal climatic periodicities, mesoscale eddies, changes in river flow, and subsidence. These local contributions, combined with extreme events that generate storm surges, primarily due to tropical and extratropical storms, result in anomalous conditions termed extreme sea level (ESL). Flood risk due to ESL is exacerbated due to its interaction with change in the trend in sea level (RSL), and hence vulnerability assessments may combine uncertainties around ESL and RSL, both in terms of contemporary assessments and future projections (e.g., Little et al., 2015b; Vousdoukas, 2016; Vousdoukas et al., 2016; Wahl et al., 2017). Changes in MSL have been dealt with in previous sections (e.g., Section 4.2.2.3.6), and here we shall focus on some of the components of ESL that have been assessed in combination with changes in MSL. Church et al. (2013) concluded that change in sea level extremes is *very likely* to be caused by a RSL increase, and that storminess and surges will contribute towards these extremes; however, it was noted that there was *low confidence* in region-specific projections.

Recent advances in statistical and dynamical modelling of wave effects at the coast (wave run up), storm surges, and inundation risk have reduced the uncertainties around the inundation risks at the coast (Vousdoukas et al., 2016) and assessments of the resulting highly resolved coastal sea levels are now emerging (Cid et al., 2017; Muis et al., 2017; Wahl et al., 2017). This progress was facilitated due to the availability of, for example, the [quasi-] Global Extreme Sea Level Analysis (GESLA-2; Woodworth et al., 2016) high-frequency dataset, advances in Coordinated Ocean Wave Climate Project (COWCLIP; Hemer et al., 2013), coastal altimetry datasets (Cipollini et al., 2017), and the Global Tide and Surge Reanalysis (GTSR; Muis et al., 2016), while new analyses of datasets (e.g., 20CR, PSMSL, DINAS-COAST; [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added]) that have been available since before the publication of AR4 or AR5 have continued [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added].

A general approach to reanalysis and projection entails coupling statistically-generated spectra of the local contributions to the components of ESL with dynamically downscaled high-resolution general circulation models (CGMs; e.g., Hemer and Trenham, 2016). Hemer and Trenham (2016) noted that wave models forced by GCMs need to be assessed for their skill in simulating historical conditions in order to determine the sources of variation between the various approaches.

Although ESL is experienced episodically, Marcos et al. (2015) examined the long-term behavior of storm surge using state space models and detected decadal and multidecadal variations in storm surge that are not related to changes in MSL. They found that, although 82% of their observed time series showed synchronous patterns at regional scales, the pattern tended to be non-linear, implying that it would be difficult to infer future behaviour unless the physical basis for the responses is understood. An analysis of the relative contributions of SLR and ESL due to storminess showed that in the US Pacific North West, increases in wave height and period have had a larger effect on coastal flooding and erosion than has had RSL (Ruggiero, 2012). This is also true in other regions (Melet et al., 2016; Melet et al., 2018). Changes in the sea level harmonics and seasonal phases and amplitudes of the wave period and significant wave height were found for the Gulf of Mexico coast since 1990 (Wahl et al., 2014; Wahl et al., 2015). They found that lower winter and higher summer sea levels have led to almost a doubling in flooding risk caused by SLR, and that the trends in the wave parameters have contributed towards an approximately 30% increase in risk of flooding. Such effects are *likely* to be highly dependent on the local conditions. For example, using WAVEWATCH III, TOPEX/Poseidon altimetry tide model data and atmospheric forcing physically downscaled using Delft3D-WAVE and Delft3D-FLOW in what they call the Coastal Storm Modeling System (CoSMoS), Barnard et al. (2014) was able to detect local hazards (at a scale of hundreds of meters) across regions along the Californian coast. Hoeke et al. (2015) showed using statistical approaches that ESL may vary by up to 1 m, over distances of less than 1 km, due to the storm track of tropical cyclones interacting with local coastal morphological properties. The addition of a 1 m RSL caused ESL to ‘modestly’ decrease, whilst resulting in the increase of wave energy impacting the coast. The finding of Hoeke et al. (2015) is typical for high oceanic islands with a narrow littoral zone that are typical of the tropics and subtropics in the Indian, Atlantic and Pacific Oceans; the failure to include wave setup when modeling inundation risk faced by such islands may lead to a significant underestimation of ESL (Hoeke et al., 2013).

A general pattern that emerges from ESL estimates at global and regional scales is that the direction of the trend in wave projections can be reasonably well modelled, but that projections of the magnitude of change are less reliable (e.g., Grabemann et al., 2015; Cid et al., 2017; Muis et al., 2017). Wahl et al. (2017) showed that for long return period events, the average combined global uncertainties around present-day ESL estimates are larger than the GMSL projection uncertainties, and at least as large as the GSML projections.

In addition to the processes above local sea level in Deltas can be dominated by subsidence. It is often a primary driver of elevated local sea level rise and increased flood hazards in those regions. This is particularly true for deltaic systems, where fertile soils, low-relief topography, freshwater access, and strategic ports have encouraged the development of many of the world's most densely populated coastlines and urban centers. It is estimated, for example, that one in fourteen of the global population resides on mid-to-low latitude deltas (Day et al., 2016). Hence, for those areas RSL is dominated by subsidence, however climate effects need to be included for estimating risks.

Deltas are formed by the accumulation of unconsolidated river born sediments and porous organic material, both of which are particularly prone to compaction. It is the compaction which results in a drop in land elevation (i.e., subsidence) that increases the rate of local sea level rise above what would be observed along a static coastline or one where only climatological forced processes control the relative sea level. Under stable deltaic conditions, the accumulation of fluvially-sourced surficial sediment and organic matter offsets this subsidence (Syvitski and Saito, 2007); however, in many cases this natural process of delta construction has been disturbed by reductions in fluvial sediment supply via upstream dams and fluvial channelization (Vörösmarty et al., 2003; Syvitski and Saito, 2007; Syvitski et al., 2009; Luo et al., 2017). Further, the extraction of groundwater, oil, and gas that fill the pore space of deltaic sediments and provide support for overlying material has significantly increased the rate of compaction and resultant subsidence along many populated deltas (Higgins, 2016).

Average subsidence rates of 6–9 mm yr^{-1} are reported for the highly populated areas of Ganges-Brahmaputra-Meghna delta in the urban centers of Kolkata and Dhaka (Brown and Nicholls, 2015). These rates will *likely* increase in the near future due to planned dam projects and an estimated 21% drop in resulting sediment supply (Tessler et al., 2018). Observations of enhanced subsidence on the Ganges-Brahmaputra-Meghna are common to most heavily populated deltaic systems. Coastal Mega-cities that have been particularly prone to human-enhanced subsidence include Bangkok, Ho Chi Minh city (Vachaud et al., 2018), Jakarta, Manila, New Orleans, West Netherlands and Shanghai (Yin et al., 2013; Cheng et al., 2018). On a global scale, observed average rates of modern deltaic subsidence range from 6–100 mm yr^{-1} (Bucx et al., 2015; Higgins, 2016). Rates of recent deltaic subsidence over the last few decades have been at least twice the 3 mm yr^{-1} rate of global mean sea level rise observed over this same interval (Higgins, 2016; Tessler et al., 2018). Numerical models that have reproduced these observed rates of deltaic subsidence by considering human-induced compaction and reduced sediment supply, support anthropogenic causes for elevated rates of subsidence (Tessler et al., 2018).

In summary, ESL and subsidence interacts with RSL change in various ways in many vulnerable areas. Therefore, we conclude with *high confidence* that the inclusion of the local processes (wave run up, storm surges, tides, erosion, sedimentation and compaction) is essential to estimate local, relative and extreme sea level changes, as in some cases they dominate over the large scale sea level rise patterns. Despite that erosion, sedimentation and compaction may be very large locally they are not accounted for in the projection sections of this chapter as no global data set are available which are consistent with RCP scenarios and the scale is often smaller than those applied in climate models.

4.2.2.6 Detection and Attribution

Attribution is the process of quantifying the evidence for a causal link between a specific external forcing—such as solar variability, volcanic eruptions, or anthropogenic changes to the atmospheric composition—and an observed change in the climate system, such as sea level change (Hegerl et al., 2010). Attribution studies can only succeed if there is understanding of the physical processes involved in translating a climate forcing into observable changes in the climate system, if an adequate representation of these processes and resulting forced change and natural variability is possible in numerical models, and if adequate observations of the investigated change of the climate system are available.

Bindoff et al. (2013) concluded that it is *very likely* that there has been a substantial contribution to ocean heat content from anthropogenic forcing since the 1970s, that it is *likely* that loss of land ice is partly caused by anthropogenic forcing, and that subsequently, it is *very likely* that there is an anthropogenic contribution to the observed trend in global mean sea level rise since 1970. However, these conclusions were based on the understanding of the responsible physical processes, instead of attribution studies dedicated to quantifying the impact of individual external forcings. Since AR5, such formal studies have attributed changes in individual contributions of sea level change (i.e., thermosteric sea level change and glacier mass loss), and in the total global mean sea level, to anthropogenic forcing.

4.2.2.6.1 Attribution of changes in individual contributions to sea level change

Marcos and Amores (2014) compared observed thermosteric sea level rise in the upper 700 m of the ocean during the period 1950–2005 with CMIP5 model reconstructions, using ‘natural-only’ forcing (i.e., solar and volcanic variability) and ‘historical’ forcing (i.e., additionally including anthropogenic greenhouse gases, aerosols, and land-use change). They used empirical orthogonal functions to maximize the signal-to-noise ratio and find that during the period 1970–2005, 87% (95% confidence interval: 72%–100%) of the observed thermosteric sea level rise is anthropogenic. (Slangen et al., 2014c) include the full ocean depth in their analysis and quantify the impact of individual forcings, by considering ‘anthropogenic-only’, ‘greenhouse gas-only’ and ‘anthropogenic aerosol-only’ CMIP5 reconstructions additionally to the ‘natural-only’ and ‘historical’ forcing CMIP5 reconstructions. They concluded that a combination of anthropogenic and natural forcing is necessary to explain the temporal evolution of observed global mean thermosteric sea level change during the period 1957 to 2005. Anthropogenic forcing was found to be responsible for the amplitude of observed thermosteric sea level change, while natural forcing was found to cause the forced variability of observations. Observations could best be reproduced by scaling the patterns from ‘natural-only’ forcing experiments by using a factor of 0.70 ± 0.30 (2 standard deviations of the CMIP5 ensemble subset used), indicating a potential overestimation of forced variability in the CMIP5 ensemble. Patterns from the anthropogenic-only’ forcing experiments needed to be scaled by a factor of 1.08 ± 0.13 (2 standard deviations of the CMIP5 ensemble subset used), indicating a realistic response of the CMIP5 ensemble to anthropogenic forcing.

Taking an approach similar to that of Marcos and Amores (2014), Marzeion et al. (2014) compared globally observed glacier mass change to results from a global glacier model forced by ‘natural-only’ and ‘historical’ reconstruction from the CMIP5 ensemble. They concluded that while natural climate forcing and long-term adjustment of the glaciers from the preceding Little Ice Age leads to continuous glacier mass loss throughout the simulation period of 1851–2010, the observed rates of glacier mass loss since 1990 can only be explained by including anthropogenic forcing. During the period 1851 to 2010, only 25 ± 35% of global glacier mass loss can be attributed to anthropogenic forcing, but since the anthropogenic fraction of mass loss is found to increase throughout the considered period, 69 ± 24% of the mass loss is attributed to anthropogenic forcing during the period 1991–2010. Uncertainties are large as glaciers are relatively small and therefore their climate conditions are poorly resolved in climate models. For ice sheet mass loss, no attempts have been made to attribute changes as time series are short with respect to the time scale and our physical understanding of ice dynamics is still too limited. However, Fyke et al. (2014) suggest that the anthropogenic signal will emerge in the surface mass balance of Greenland within the first half of the 21st century (increased melt in the periphery, and increased accumulation in the center). The effects of groundwater depletion and reservoir impoundment on sea level change are anthropogenic by definition (e.g., Wada et al., 2012).

4.2.2.6.2 Attribution of global mean sea level change

By estimating a probabilistic upper range of long-term persistent natural sea level variability, Dangendorf et al. (2015) attributed the fraction of observed sea level change unexplained by natural variability to anthropogenic forcing and concluded by inference that it is virtually certain that at least 45% of the observed increase in global mean sea level since 1900 is attributable to anthropogenic forcing. Similarly, Becker et al. (2014) provided statistical evidence that the observed sea level trend, both in the global mean and at selected tide gauge locations, is not consistent with unforced, internal variability. They concluded by inference that more than half of the observed global mean sea level trend during the 20th century is attributable to anthropogenic forcing.

Using a semi-empirical model relating rates of sea level change to global mean temperature anomalies and counter-factual temperature scenarios for the 20th century, Kopp et al. (2016) concluded that it is *extremely likely* that 49% of the observed global mean sea level rise during the 20th century is attributable to anthropogenic forcing.

Slangen et al. (2016) reconstructed global mean sea level from 1900 to 2005 based on CMIP5 model simulations including 'natural-only', 'greenhouse gases-only', 'anthropogenic aerosols-only', 'anthropogenic-only' and 'historical' forcing and combining the contributions of thermosteric sea level change with glacier and ice sheet mass loss. They found that the naturally caused sea level change, including the long-term adjustment of sea level to climate change preceding 1900, caused 67 ± 23% of observed change from 1900 to 1950, but only 9 ± 18% between 1970 and 2005. Anthropogenic forcing was found to have caused 15 ± 55% of observed sea level change during 1900–1950, but 69 ± 31% during 1970 to 2005. However, the sum of all contributions explains only 74 ± 22% of observed global mean sea level change during the period 1900–2005 (considering the mean of the reconstructions of Church and White, 2011; Ray and Douglas, 2011; Jevrejeva et al., 2014; Hay et al., 2015; implying that a minor fraction of observed sea level change remains unattributed.)

Based on these multiple lines of evidence, we conclude with *high confidence* that anthropogenic forcing *very likely* is the dominant cause of observed global mean sea level rise since 1970.

4.2.2.6.3 Regional detection, local emergence of an anthropogenic signal

Since variability of sea level on the regional scale is larger than on the global scale, attribution of observed regional sea level change to specific climate forcings is more challenging. For example, Palanisamy et al. (2015b) show explicitly that the anthropogenic fingerprint of sea level rise predicted by CMIP5 models for the Pacific Ocean is too small to be detected in altimetric observations, compared to the level of internal variability, in the considered region mostly associated with the Pacific Decadal. In several regions climate models are not able to reproduce the unforced and forced signal in sea level. Sérazin et al. (2016) show limitations for western boundary currents and Bilbao et al. (2015) discuss shortcomings in the Pacific Ocean.

In a related approach, a number of studies have addressed the observation period necessary to locally detect an anthropogenic signal in sea level rise. Lyu et al. (2014) concluded that relative to the reference period 1986 to 2005, the anthropogenic signal in sea level change will be detectable in 50% of the ocean area by 2020. Similarly, Richter and Marzeion (2014) concluded that relative to 1990, the forced signal will become detectable by 2020 locally in ocean areas with low internal variability of sea level, such as the tropical Atlantic Ocean, where also Bilbao et al. (2015) predict the earliest detectability of an anthropogenic trend. Jordà (2014) showed that the time needed for local detection of a centennial linear trend of 2 mm yr^{-1} is on average 40 years. Richter et al. (2017) showed that the glacier contribution increases the detectability of sea level change away from the locations of ice mass loss, and that spatial smoothing on a scale of 2000 km leads to detectability of an anthropogenic signal in 93% of the ocean area when considering the period 1970–2015.

We conclude with *medium confidence* that an attribution of regional and local mean sea level change is not yet possible.

4.2.2.6.4 Attribution of sea level extremes

The conventional approach of attribution to external climate forcing is not applicable to individual extreme events, which are unique by definition and whose occurrence is strongly influenced by chance (Trenberth et al., 2015). However, it is possible to quantify the evidence that anthropogenic climate change has altered the probability of a type of event occurring (e.g.,Trenberth et al., 2015; Otto et al., 2016; Stott, 2016). For example, Takayabu et al. (2015) find evidence that the storm surge of Typhoon Hayan was intensified through anthropogenic influence. They find that removing the anthropogenic warming signal of the sea surface and the atmosphere leads to a decrease in simulated wind speeds and an increase in simulated core pressure in 15 out of 16 ensemble members, increasing the differences between simulated and observed values. Removing the anthropogenic signal further leads to a mean decrease of the storm surge height of around 20%.

4.2.3 Projections of Sea Level Change

As a consequence of climate change global and regional mean sea level will change. Hence, we use AOGCM models are used to make projections of the climate changes and the associated sea level rise. Atmosphere Ocean General Circulation Models (AOGCMs) can be applied on century time scales, to provide estimates of the steric (temperature and salinity effects on sea water density) and ocean dynamical (ocean circulation) components of sea level change, both globally and regionally. GCMs also resolve climate variability related to changes in precipitation and evaporation, which is relevant for changes in the hydrological cycle which play a role in shorter duration sea level changes (Cazenave et al., 2014; Hamlington et al., 2017). With various degrees of success those models capture El Niño-Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO) and other modes of variability which affect sea level through redistributions of energy and salt in the ocean. Results from the CMIP5 AOGCM archive produced for AR5, are used to provide information on expected changes in the oceans, and evolving climate glaciers and ice sheets. New estimates from CMIP6 are not yet available and will be discussed in AR6. These models do not (yet) explicitly calculate changes in ice mass, partly because of the small scales to resolve ice sheet processes properly, and partly because the relevant physical processes are poorly understood. Typically, ice sheet and glacier changes are calculated based on the relevant variables of ocean and atmospheric temperature and precipitation. These off-line climatologies can be dynamically or statistically downscaled to match the high temporal resolution required for ice sheets and glaciers, but serious limitations remain. Particularly problematic is the lack of interactive coupling between the land ice, ocean, and atmospheric components. This deficiency prohibits adequate representation of potentially important feedbacks between changes in ice sheet geometry and climate, for example through fresh water and iceberg impacts on ocean circulation and sea ice, which can have global consequences (e.g., Lenaerts et al., 2016). Dynamics of the interaction of ice streams with bedrock and till at the ice base remains difficult to model due to lack of supporting observations. Nevertheless, many new ice sheet models have been generated over the last few years, particularly for Antarctica (Section 4.2.3.1) focusing on the dynamic contribution of the ice sheet to sea level change, which remains the key uncertainty in future projections (Church et al., 2013), particularly beyond 2050 (Nauels et al., 2017a; Slangen et al., 2017b; Horton et al., In Press).

Information beyond that provided by climate models is needed to describe local and relative sea level changes. Geodynamic models are used to calculate relative sea level changes due to mass changes in past and future. This includes both Earth gravitational and rotational changes, as ice and water are redistributed around the globe, and glacial isostatic adjustment (GIA). Input for those models are provided by the mass changes following from the off-line ice models, time series of terrestrial water mass changes which typically require climate input, and reconstruction of past ice sheet changes over the last glacial cycle. Combining different models leads to projections of RSL (Section 4.2.3.2).

At the local spatial scales of specific cities, islands, and stretches of coastlines, the impacts of highly variable processes leading to ESL, such as tropical cyclone-driven storm surges, hydrodynamical models are required (Section 4.2.3.3) as well as knowledge on sedimentation and erosion. These models are capable of providing statistics on the variability or the change in variability of the water level required for flood risk calculations at specific locations and at spatial scales of less than 1 km. The models also rely on input from climate models, like temperature, precipitation, and wind regime, and storm tracks (Colbert et al., 2013; Garner et al., 2017).

In summary, climate models play an important role, in addition to emission scenarios, geodynamic, ice-dynamic, and hydrodynamic models, during the various steps of projections, to provide required information for hazard estimation for coasts and low-lying islands. In this report we rely on results of the CMIP5 model runs.

Besides this sequence of sections leading to projections of ESL, we address the uncertainties and decadal predictability of sea level (Section 4.2.3.4) and the long-term scenarios, beyond 2100 (Section 4.2.3.5).

4.2.3.1 Dynamic Contribution of Ice Sheets

4.2.3.1.1 Greenland

The Greenland Ice Sheet (GIS) is currently losing mass at roughly twice the pace of the Antarctic Ice Sheet (AIS; Hanna et al., 2013; Csatho et al., 2014; Enderlin, 2014; McMillan et al., 2016; van den Broeke, 2016). Ice loss on Greenland, equivalent to ~0.47 ± 0.23 mm yr^{-1} of GMSL rise averaged over 1991–2015, has been dominated (60%) by increasingly negative surface mass balance (SMB) caused by surface melt and runoff on the lower elevations of the ice sheet's margins, rather than ice dynamical changes (Csatho et al., 2014; Enderlin, 2014; van den Broeke, 2016). Recent ice sheet modelling (Edwards et al., 2014; Fürst et al., 2015; Vizcaino et al., 2015) indicates this trend of increasing surface melt will dominate Greenland's contribution to GMSL throughout the 21st-century, regardless of which emissions scenario is followed (*medium confidence*). Hence, ice dynamical changes are thought to be less important for Greenland. One reason for this is the limited volume of ice with direct access to the ocean, as shown in Figure 4.5, which illustrates a fundamental geometrical difference between Greenland and Antarctica. In Greenland, most of the bedrock at the ice-sheet margin is above sea level. The opposite condition exists in Antarctica and in places where the subglacial bedrock slopes downward, away from the coast (reverse-sloped), the glacial ice is susceptible to dynamical instabilities that can contribute rapid ice loss to the ocean.

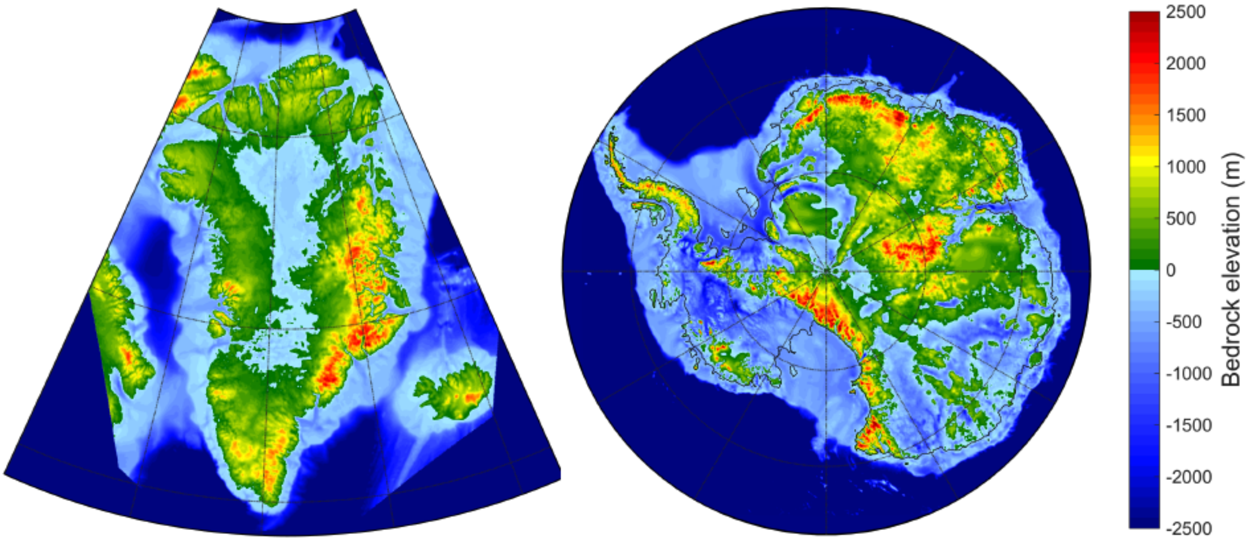


Figure 4.5: Bedrock topography below the existing ice sheets in Greenland and Antarctica (Fretwell et al., 2013; Morlighem et al., 2017). The thin black line on the Antarctic map shows the location of the present day grounding line, which is well below sea level around much of the continent. Mind that the horizontal scales are not the same in both panels [PLACEHOLDER FOR SECOND ORDER DRAFT: scale in km to be added].

Fürst et al. (2015) used ten different CMIP5 AOGCM simulations to provide offline SMB and ocean forcing for a Greenland-wide ice sheet model, accounting for influences of warming subsurface ocean temperatures and basal lubrication on ice dynamics. In their RCP8.5 ensemble, they found a GIS contribution to GMSL ranging from 5.1 to 16.6 cm (10.15 ± 3.24). This wide range of RCP8.5 results highlights substantial climate-driven uncertainty in 21st-century GIS projections, as found in other studies (Edwards et al., 2014). These results support the previous conclusions of Shannon et al. (2013), showing that dynamical changes caused by basal meltwater and lubrication have a limited impact on the rate of mass loss. The median estimates of Fürst et al. (2015) are in good agreement with previous multi-model results (Bindschadler et al., 2013) and the assessment of AR5 (Church et al., 2013), which reported a *likely* RCP8.5 range of Greenland's contribution to GMSL between 7 and 21 cm by 2100.

Detailed flowline modeling of four Greenland outlet glaciers (Petermann, Kangerdlugssuaq, Jakobshavn Isbræ, and Helheim) (Nick et al., 2013) reported a dynamical contribution to sea level in an RCP8.5 scenario of 8.5–13.1 mm, although their study did not include the Northeast Greenland Ice Stream (NEGIS), which has recently been reported to be a potentially important contributor to future sea level (Khan et al., 2014). In their Greenland-wide model, Fürst et al. (2015) found an overall reduction in the rate of dynamic ice discharge to the ocean. This response, caused by ongoing thinning of the ice sheet margin and landward retreat of outlet glaciers away from the coast, suggests Greenland's potential for a rapid, dynamic contribution to sea level may be topographically 'self-limiting', as found in other ice-modeling studies (Goelzer et al., 2013; Lipscomb et al., 2013; Vizcaino et al., 2015). Importantly, recent subglacial mapping since AR5 has uncovered extensive, deep valley networks extending into the GIS interior (Morlighem et al., 2014; Morlighem et al., 2017). These new data show that the termini of many marine-terminating outlet glaciers are deeper than previously known and may be more exposed to warm subsurface Atlantic Water than previously considered. While detailed and accurate subglacial topography has been shown to be important in the modeling of Greenland outlet glaciers (Aschwanden et al., 2016), the potential for these newly revised bedrock boundary conditions to substantially change the outcome of Greenland-wide model simulations is not yet known.

In summary, new model guidance appearing since AR5 (Fürst et al., 2015; Vizcaino et al., 2015) builds on previous studies suggesting future Greenland ice loss will be dominated by surface processes, rather than dynamic ice discharge to the ocean (*medium confidence*). Based on these modeling studies, GIS is not expected to contribute more than 20 cm of GMSL rise by 2100 in a RCP8.5 scenario, similar to the upper end of the *likely* range reported by AR5 (Church et al., 2013). Confidence in the projections will remain low until the dynamical implications of newly discovered subglacial topographic features and bathymetric details at the ice margin (Morlighem et al., 2017) are more thoroughly tested. Different poorly understood physical mechanisms play a role in those narrow fjord regions than in the much wider region in Antarctica, like Thwaites. Greenland ice-sheet simulations are sensitive to uncertainties in the applied climate forcing (Edwards et al., 2014), but updated climate projections since AR5 are not yet available. Because of the consistency of recent modeling studies with the assessment of Church et al. (2013), and lack of updated climate guidance, we use Greenland's contribution to future sea level reported in AR5 in our projections of GMSL.

4.2.3.1.2 Antarctica

The Antarctic Ice Sheet (AIS) contains almost eight times more glacial ice above flotation (grounded ice above sea level that causes sea level rise if lost to the ocean) than Greenland. One third of the ice sheet sits on bedrock hundreds of meters (or more) below sea level (Fretwell et al., 2013) and most of its margin is in direct contact with the ocean (Figure 4.5). These geographic features make the overlying ice sheet vulnerable to dynamical instabilities that can cause rapid ice loss (Weertman, 1974; Schoof, 2007a; Pollard et al., 2015). Changes in both the surrounding ocean affecting sub-ice oceanic melt rates, and changes in the overlying atmosphere affecting surface mass balance and surface meltwater production can trigger these instabilities, but the timing, magnitude, and potential pace of future retreat remains deeply uncertain.

In contrast to Greenland, Antarctica's recent contribution to sea level rise (Helm et al., 2014; Velicogna et al., 2014; Williams et al., 2014; Martín-Español et al., 2016; Martin-Español et al., 2017) has been dominated by ice-dynamical processes (Helm et al., 2014; Mouginot et al., 2014; Rignot et al., 2014; Velicogna et al., 2014; Williams et al., 2014; Li et al., 2015b; Khazendar et al., 2016; Martín-Español et al., 2016; Scheuchl et al., 2016; Martin-Español et al., 2017; Seroussi et al., 2017), rather than changes in surface mass balance. This mass loss is concentrated in the Amundsen Sea and Bellingshausen Sea sectors of the West Antarctic Ice Sheet (WAIS), where the termini of outlet glaciers are in direct contact with the ocean. Since AR5, it has become increasingly evident that ice loss in this region is being driven by sub-ice oceanic melt (thinning) of ice shelves (Paolo et al., 2015; Wouters et al., 2015; Khazendar et al., 2016) and the resulting loss of backpressure (buttressing) that impedes the seaward flow of grounded ice upstream. Glacier retreat, increasing most where local ocean warming is greatest, provides additional confidence that the AIS is currently responding to oceanic warming (Paolo et al., 2015; Khazendar et al., 2016). In the Amundsen Sea, the subsurface ocean warming is driven by localized upwelling of warm circumpolar deep waters (CDW; (Schmidtko et al., 2014; Khazendar et al., 2016), which points to the potential for ongoing ocean warming and ocean heat uptake to be impactful to the extent that it affects the ocean near the ice fronts. Whether this process is anthropogenically influenced or part of the regional variability in ocean circulation is unclear.

Where grounding lines are poised on reverse sloped bedrock, the initial thinning of a marine terminating ice margin (possibly initiated by the thinning or loss of buttressing ice shelves), has long been theorized to have the potential to trigger a positive feedback resulting in rapid ice-sheet retreat (Weertman, 1974). This so called 'Marine Ice Sheet Instability' (MISI), described extensively in AR5, operates as a positive feedback, because the seaward flow of ice at the grounding line is strongly dependent on its vertical thickness (Schoof, 2007a). As a consequence, if retreat is initiated on a reverse-sloped bed, the grounding line thickness will continue to increase, as will the seaward flux of ice. The grounding line might continue to retreat until it re-stabilizes on a topographic feature, or if adequate back stress is supplied by a confined ice shelf, lateral shear, or some other mechanism. As such, the onset and persistence of MISI is dependent on several factors in addition to bed slope, including the details of the bed geometry, width of channelized flow, basal traction, side sheer, self-gravitation effects on local sea level at the grounding line, and ice-shelf pinning points. Hence, long-term retreat on every reverse-sloped bed is not necessarily unstoppable (Gudmundsson et al., 2012; Parizek et al., 2013; Docquier et al., 2014; Gomez et al., 2015), however, there is growing observational and modeling evidence since AR5 that MISI-style retreat is indeed underway in several major Amundsen Sea outlets, including Thwaites, Pine Island, Smith, and Kohler Glaciers (Favier et al., 2014; Mouginot et al., 2014; Rignot et al., 2014; Seroussi et al., 2017).

Totten Glacier, the largest outlet in East Antarctica and draining a deep submarine basin containing enough ice to raise GMSL by ~3.9 m, has also been retreating and thinning in recent decades (Li et al., 2015b). Like the Amundsen Sea outlets, Totten's retreat has been connected to localized oceanic warming and sub-ice melting of its thinning ice tongue (Khazendar et al., 2013; Li et al., 2016; Rintoul et al., 2016; Greene et al., 2017). The oceanic warming has been attributed to an increase in local polynya activity (Khazendar et al., 2013) and wind-driven incursion of warm, salty deep water in deep submarine channels (Spence et al., 2014; Rintoul et al., 2016). Totten's recent behaviour points to the possibility that East Antarctica could become a substantial contributor to future sea level rise. Geological and modeling evidence indicate repeated retreat and re-advances of the glacier over the last few million years, particularly during the warmth of the Pliocene (Aitken et al., 2016), but the implications for its response to future warming are largely unknown.

A number of ice-sheet modeling studies since AR5 have focused on the potential response of WAIS to increasing sub-ice shelf and grounding zone melt rates (e.g., Cornford et al., 2015; Feldmann and Levermann, 2015; Arthern and Williams, 2017). Sub-ice melt rates are highly sensitive to ocean temperature (quadratic dependency), water depth, boundary layer and turbulent processes at the ice-ocean interface, as well as the local ice shelf cavity geometry (Jenkins, 1991; Holland et al., 2008). Progress has been made to include these processes in sub-ice melt models (Schodlok et al., 2016; Reese et al., 2017; Lazeroms et al., 2018), but sub-ice melt continues to be crudely parameterized in most continental scale ice dynamical models, typically only capturing the temperature and depth-dependence on melt potential. Such simplified parameterizations lacking two-way interaction between a retreating ice margin and the surrounding ocean can overestimate melt rates in time-evolving simulations of marine based glacier retreat (Seroussi et al., 2017).

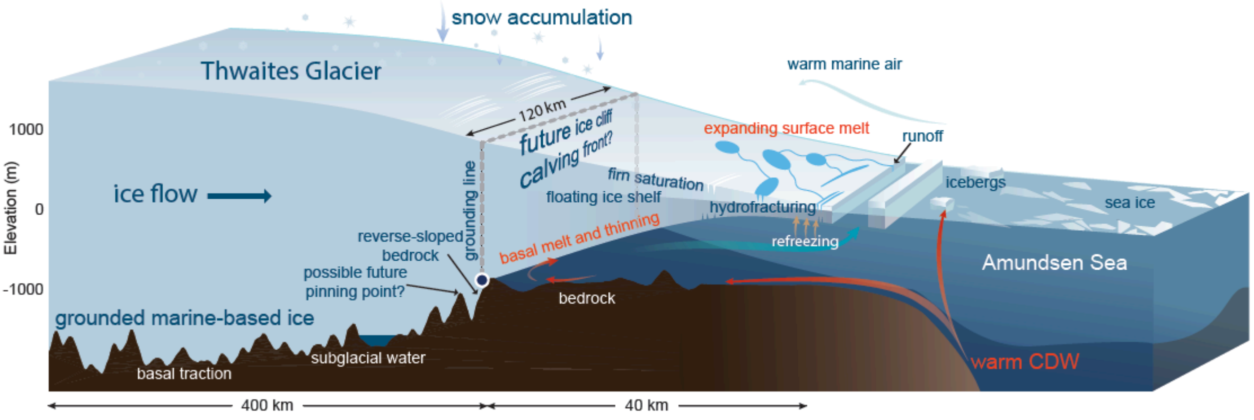


Figure 4.6: Processes affecting the Thwaites Glacier in the Amundsen Sea sector of Antarctica. The grounding line is currently retreating on reverse-sloped bedrock at a water depth of ~600m (Joughin et al., 2014; Mouginot et al., 2014).

The glacier is 120 km wide, widens upstream, and is minimally buttressed by a laterally discontinuous ~40 km long ice shelf. The remaining shelf is thinning in response to warm, sub-shelf incursions of circumpolar deep water (CDW), with melt rates up 60 m yr^{-1} near the groundling line (Rignot et al., 2014; Schodlok et al., 2016; Seroussi et al., 2017). The bathymetry upstream of the grounding zone is complex, but it generally slopes downward into a deep basin, up to 2,000 m below sea level under the center of the WAIS (far left). By itself, Thwaites contains enough ice to raise GMSL by ~0.4 m (Holt et al., 2006; Millan et al., 2017), but it could have a destabilizing impact on the broader WAIS (Feldmann and Levermann, 2015), which contains the equivalent of > 3.2 m of sea level rise. Atmospheric processes and surface meltwater may soon begin to play an increasingly important role in addition to the ocean-driven retreat already underway (Scambos et al., 2017).

Studies using highly resolved (a few km or less) ice models have mostly been limited to the study of single outlet glaciers (Favier et al., 2014; Joughin et al., 2014; Seroussi et al., 2017), or to WAIS only (Cornford et al., 2015; Nias et al., 2016). While limited to 50-yr simulations, Seroussi et al. (2017) provide the first, interactively coupled ice-ocean model simulations of Thwaites Glacier at a high spatial resolution. Like Joughin et al. (2014), their model demonstrates MISI-like grounding line retreat at a rate of ~1 km yr^{-1}, comparable to observations between 1992 and 2011 (Rignot et al., 2014). However, the retreat is interrupted when the main trunk of the glacier stabilizes on a bathymetric ridge, ~20 km upstream of the present-day grounding line (Fig. 4.5). Due to the short duration of the simulation, the long-term potential for retreat into the interior of the ice sheet is not captured. Cornford et al. (2015) used a dynamical ice sheet model with an adaptive mesh that maintains very high spatial resolution at the grounding zone. This represents a significant modeling advance relative to most studies before AR5. In an idealized, extreme ocean warming scenario, they demonstrate that rapid ice-shelf thinning can produce up to 20 cm of GMSL rise from WAIS alone by 2100. However, using more realistic climate and ocean forcing representing an A1B scenario, they find only 5 cm of GMSL rise by 2100, some of which is compensated by increased precipitation over the ice sheet. Similarly to Seroussi et al. (2017), they find strong dependency of Thwaites Glacier retreat on model resolution, initial conditions, and surface mass balance forcing. Arthern and Williams (2015) also use adaptive mesh techniques, but with a different (vertically integrated) formulation to simulate the response of the Amundsen Sea sector of West Antarctica to increasing sub-ice shelf melt rates. Their multi-century simulations without increased melting support the notion that sustained long-term retreat of the Amundsen Sea outlets is already underway. They use two alternative parameterizations of sub-ice shelf melting. The first applies melt only to the bottom of fully floating grid cells, while the second uses a sub-grid parameterization that applies some melt to partially grounded grid cells. The later treatment substantially increases the response of the ice sheet to oceanic melting, pointing to the need for additional analysis of marine melt-rate parameterizations and their potential to introduce model-dependent behaviour. In an ensemble of simulations using a range of model physical parameters, Nias et al. (2016) demonstrated substantial model sensitivity to poorly resolved basal boundary conditions (topography and basal traction), also contributing to model uncertainty. Ongoing uncertainty in future Thwaites retreat is critical, because the 120 km wide glacier is poised on a mostly reverse sloped bed (Millan et al., 2017), that reaches more than 300 km upstream into the heart of the WAIS where the ice is up to 2 km thick (Fig. 4; (Scambos et al., 2017). The WAIS contains enough ice above floatation to cause > 3.2 m of GMSL rise (Bamber et al., 2009) if lost to the ocean.

In summary, these targeted modeling studies demonstrate the potential for warming ocean temperatures to initiate grounding line retreat and thinning of upstream ice through reduced buttressing. This suggests the onset of MISI in some Amundsen Sea outlets is already underway, as supported by observations of thinning and grounding line retreat (Mouginot et al., 2014; Rignot et al., 2014), including the Thwaites Glacier which penetrates the interior of the West Antarctic Ice Sheet (Figure 4.6). However, the irreversibility of retreat and the long-term implications for the wider WAIS remain uncertain.

Since AR5, atmospheric forcing has become increasingly recognized to be an important factor for the future of the AIS, as it is on Greenland today. A sustained (15 days) melt event over the Ross Sea sector of the WAIS in 2016, illustrated both the connectivity of Antarctica to the tropics and El Niño, and the possibility that future meltwater production on ice shelf surfaces could fundamentally change in the near future (Nicolas et al., 2017). This was highlighted by Trusel et al. (2015), who used the RACMO2 regional atmospheric model under RCP8.5 (Kuipers Munneke et al., 2012), to demonstrate a substantial expansion of surface meltwater production on ice shelf surfaces after 2050 that exceed melt rates observed before the 2002 collapse of the Larsen B Ice Shelf. Surface meltwater is important for both ice sheet dynamics and surface mass balance due its potential to lower albedo, saturate the firn layer, deepen surface crevasses, and to cause flexural stresses that can contribute to ice shelf break up (hydrofracturing; Banwell et al., 2013; Kuipers Munneke et al., 2014). When and if melt rates will be sufficiently high in future warming scenarios to trigger widespread hydrofracturing is still under debate.

Continental-scale ice sheet simulations are ultimately required to provide projections of future GMSL rise from Antarctica. However, due to the spatial scale of the region, and complex interactions between the atmosphere, ocean, sea ice, ice shelves and ice sheets, existing model simulations have yet to include these interacting systems collectively. They also rely on simplifying approximations of the equations representing three-dimensional ice flow, and in some cases, they parameterize ice flow at the grounding line (Schoof, 2007a) to improve computational efficiency. Such simplifications are necessary to allow long simulations that be validated against geological information, in addition to modern observations (Briggs et al., 2013; Pollard et al., 2016). Processes related to MISI are best represented at high spatial resolution and without simplifications of the underlying physics (Favier et al., 2014; Cornford et al., 2015). However recent model intercomparisons have shown simplified, continental-scale models can perform reasonably well relative to highly resolved models with more explicit physical treatments, and can capture grounding line dynamics and the essence of MISI (Pattyn et al., 2012; Pattyn and Durand, 2013). Accurate atmospheric forcing (SMB) and sub-ice melt (Arthern and Williams, 2015; Golledge et al., 2015) are also crucial prerequisite to resolving the time-evolving dynamics of the system.

Since AR5, several Antarctic continental-scale models have been applied to future greenhouse gas scenarios on century and longer timescales (Levermann et al., 2014; Golledge et al., 2015; Ritz et al., 2015; Winkelmann et al., 2015; Clark et al., 2016; DeConto and Pollard, 2016). Ritz et al. (2015) used a hybrid physical-statistical modeling approach, whereby the timing of MISI onset is determined statistically rather than physically. They estimated probabilities of MISI onset in eleven different sectors around the ice-sheet margin based on observations of Amundsen Sea retreat over the last few decades, and expected future climate change following an A1B emission scenario only. In places where MISI is projected to begin, the persistence and rate of grounding-line retreat is parameterized as a function of the local bedrock topography (slope), grounding line thickness following Schoof (2007a), and a formulation for basal friction. The advantage of this approach is that the relative simplicity of the ice model allows thousands of iterations, allowing a probabilistic assessment of the results and the calibration with present-day retreat rates. While their A1B future climate scenario is not directly comparable to the RCPs used in other studies, they concluded that Antarctica could contribute up to 30 cm GMSL by 2100 (95% quantile, nearly Gaussian distribution). This study represents a statistically rigorous approach in which model parameters are based on a synthesis of observations and projected surface and sub-shelf forcing, rather than from climate and ocean models. However, the model calibrations rely on recent observations, which may not provide adequate future guidance under warmer climate and ocean conditions. In addition, their model considers only processes associated with MISI, and does not consider possible contributions from other physical processes which may emerge, for which no recent analogue exists. Relatively little attention is given to the role of the SMB or BMB.

Golledge et al. (2015) used the PISM ice sheet model (Winkelmann et al., 2011), to simulate the future response of the AIS to RCP emission scenarios. They did not attempt to calibrate their model to the current observations as in Ritz et al. (2015). The PISM model links grounded, streaming, and shelf flow, has freely evolving grounding lines, and captures MISI dynamics. PISM's parameterized treatment of sub-ice melt in response to warming ocean temperatures (Feldmann and Levermann, 2015) makes the model sensitive to subsurface ocean warming, which is extrapolated from surface ocean temperatures simulated by a simple slab ocean model. They simulated a 39 cm contribution to GMSL by 2100 in RCP8.5, mainly through MISI, but using a more conservative oceanic melt-rate parameterization, the GMSL contribution is reduced from 39 to 10 cm. This difference highlights the ongoing uncertainty in heavily parameterized continental-scale ice sheet models, including their sensitivity to ocean forcing. While providing alternative outcomes with the two basal melt rate parameterizations, they do not provide a probability distribution for their results.

Levermann et al. (2014) used simplified emulations of temperature increase in order to calculate both sub-ice melt and SMB to calculate the dynamic response for five ice-sheet models calibrated against recent rates of retreat and including a parameterized delay for ocean warming. For scenarios including a delay in ocean warming they find 0.0–0.23 m GMSL for RCP2.6 (90% range) and 0.01–0.37 for RCP8.5 in 2100.

Substantial uncertainty arises from the different model treatments of grounding line dynamics and ice shelves, however they conclude that the single greatest source uncertainty stems from the external forcing.

DeConto and Pollard (2016) used an ice sheet model with a formulation similar to that used by Golledge et al. (2015), but they include two fundamental glaciological processes not accounted for in other continental scale models: 1) surface melt and rain water influence on crevasse penetration in divergent flow regimes (hydrofracturing); and 2) structural failure of marine- terminating ice fronts that have lost their ice shelves due to ocean melt and hydrofracturing, and are tall enough (~800 m) to generate stresses that exceed the strength of the ice (Bassis and Walker, 2012). These hydrofracturing and marine ice cliff instability processes (Box 4.1) are represented with simplistic parameterizations, but their inclusion improves the model's ability to match albeit uncertain geological sea level targets in the Pliocene (Pollard et al., 2015) and Last Interglacial (DeConto and Pollard, 2016). Most Pliocene sea level estimates imply ice sheet retreat into deep East Antarctic basins, in addition to the loss of the Greenland and West Antarctic ice (Dutton et al., 2015). Mechanisms other than ice cliff collapse have been hypothesized that could drive substantial East Antarctic ice loss in the absence of buttressing ice shelves. As such, the MICI solution to the Pliocene sea level problem (Pollard et al., 2015) may not be unique (Aitken et al., 2016). For example, a recent ice sheet modeling study (Pattyn, 2017) found that using a basal sliding scheme based on Coulomb friction near the grounding line (Tsai et al., 2015) leads to the rapid loss of the WAIS and major retreat in East Antarctic basins if all buttressing ice shelves are suddenly removed from the model. The end result is similar to Pliocene simulations with hydrofracturing of ice shelves and MICI (Pollard et al., 2015), although the complete removal of all ice shelves in the Coulomb friction study may not be physically justifiable and is not relatable to a Pliocene scenario with modest warming. Further justification for developing hydrofracture and ice-cliff calving parameterizations include the observed break up of ice shelves in response to surface meltwater Scambos et al. (2004); Scambos et al. (2009) and direct observations of ice-cliff failure in the few places where thick (> 800 m) marine terminating grounding lines have lost their buttressing ice shelves (e.g., Jakobshavn, Helheim, and Crane glaciers). Further support for ice cliff failure to contribute to rapid ice loss is provided by a modelling study by Parizek et al. (Submitted), who argue that retrogressive slumping caused by the stress differences near the cliffs are the key instability mechanism.

Inclusion of hydrofracturing and MICI processes, substantially increases projected Antarctic contributions to GMSL in RCP4.5 and RCP8.5 ensembles, driven by offline regional atmospheric and ocean model climatologies. DeConto and Pollard (2016) provide four alternative ensembles for each RCP scenario, representing two alternative ocean model treatments (with and without an ocean temperature bias correction), and two alternative Pliocene sea level targets used to tune their model physics. The model ensembles use a range of uncertain model parameters associated with hydrofracturing and MICI, validated relative to Last Interglacial and Pliocene sea level targets, however; their simulations do not explore the full range of model parameter space, and their simple statistical treatment of ensemble results don't provide a probabilistic assessment of Antarctica's future (Horton et al., In Press). Their four RCP4.5 and RCP8.5 ensemble means range between 0.26–0.58 m and 0.64–1.14 m, respectively by 2100. RCP8.5 is shown to produce as much as 15 m of GMSL sea level rise by 2500, mainly from the retreat of ice in deep East Antarctic basins in addition to West Antarctica. Golledge et al. (2015) and DeConto and Pollard (2016) find very little GMSL rise from Antarctica in their RCP2.6 scenario (0.02–0.16 m), implying a much reduced probability of extreme sea level rise from Antarctica under strong mitigation. A few individual RCP2.6 ensemble members simulate up to 0.5 m of sea level rise by 2100, mainly through the rapid retreat of Thwaites Glacier, reinforcing the ongoing uncertain sensitivity of this major outlet glacier to warming reported in other studies (Cornford et al., 2015; Nias et al., 2016; Seroussi et al., 2017). DeConto and Pollard (2016) study lacks quantitative calibration with present-day retreat rates, there is large uncertainty in their SMB model and the onset of surface meltwater production, and sub-ice melt, but it does point to the potential for physical processes not considered by AR5 to be strongly impactful on future rates of GMSL rise.

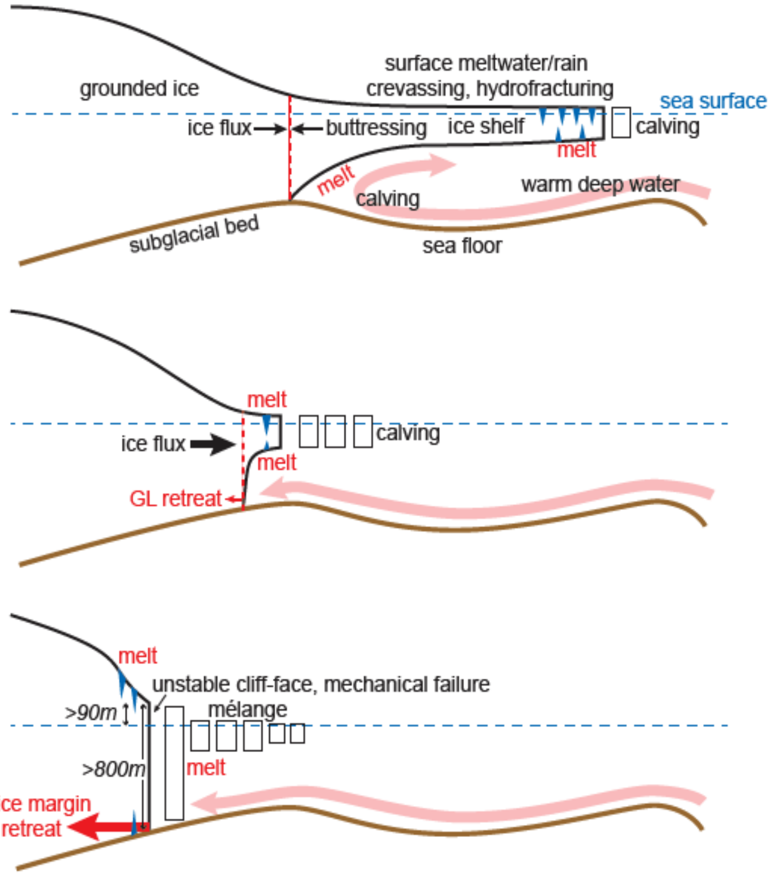
[START BOX 4.1 HERE]

Box 4.1: Recent Advances in Ice Sheet Models

The seaward flow of ice at the grounding line increases non-linearly with respect to the thickness of the ice at the grounding line (Schoof, 2007a). As a result, more and more ice will flow into the ocean as the ice margin backs onto a retrograde bed. This is the essence of the marine ice sheet instability (MISI). This increase in seaward ice flow can reform buttressing ice shelves, possibly stalling or halting MISI and MICI, despite a warming ocean. However, under high emissions scenarios, Antarctic ice shelves are projected to become flooded with surface rain and melt water (Trusel et al., 2015), with the potential for ongoing ice-shelf breakup through hydrofracturing. In this case, tall calving ice cliffs would persist, despite the seaward increase in ice flow. In sum, hydrofracturing and ice-cliff collapse operate collectively to produce MICI.

The meltwater-induced loss of ice shelves and onset of widespread MICI in Antarctica would substantially increase the pace of sea level rise through the calving of the cliffs into the ocean, in addition to ice-dynamical processes associated with MISI. With the exception of Crane Glacier, marine-terminating ice in Antarctica with ice at grounding lines thick enough to produce unstable ice cliffs are currently protected by buttressing ice shelves, so the potential for MICI to contribute to sea level rise remains largely hypothetical. Possible geophysical evidence of past MICI in Antarctica is provided by deep iceberg plough marks on the sea-floor (Wise et al., 2017), however this interpretation based on a single study is speculative. MICI is based on fundamental force balance calculations and physical principles and its inclusion in a continental ice sheet model facilitates the simulation of Pliocene and LIG sea level high-stands (DeConto and Pollard, 2016; Pollard et al., 2016). However, MICI involves small-scale processes related to brittle fracture mechanics, and it has only been represented by simple parameterizations in one large scale ice-sheet models to date. While MICI has the potential to be highly impactful for future sea level, it remains deeply uncertain.

[PLACEHOLDER FOR SECOND ORDER DRAFT: to be expanded with Adaptive Mesh Refinement and coupled ice-ocean simulations]



Box 4.1, Figure 1: Schematic representation of the marine ice cliff instability (MICI) hypothesis, adapted from Pollard et al. (2015) and DeConto and Pollard (2016). A combination of sub-ice oceanic melting and surface meltwater-induced hydrofracturing leads to ice shelf loss (top and middle). Where the marine terminating margin is thick enough to expose tall ice cliffs ~100 m above sea level, the stresses at the cliff face exceed the strength of the ice (Bassis and Walker, 2012; National Centers for Environmental Information, 2017) and the cliff face fails structurally in repeated calving events (bottom), similar to the behavior seen today at the termini of Greenland's Jakobshavn and Helheim glaciers, and

at Crane glacier on the Antarctic Peninsula after the collapse of the Larsen B ice shelf. The loss of buttressing ice shelves can initiate thinning and retreat of the ice sheet margin, with or without ice cliffs (middle panel).

[END BOX 4.1 HERE]

While these simulations point to the potential for a far greater contribution to sea level than other studies, particularly on longer time scales, deep uncertainty remains. Accounting for the influence of surface meltwater on ice shelf breakup (hydrofracturing) makes the timing of retreat particularly sensitive to the emergence of daily summer temperatures above 0 ºC. In this case, SMB is determined by a single atmospheric model that produces more melt, earlier in the 21st century than the well validated snow/firn model used by Trusel et al. (2015). Realistically capturing the meltwater-buffering capacity of the firn layer is important, because saturated, meltwater has the potential to flow into underlying crevasses to cause hydrofracturing (Kuipers Munneke et al., 2014). Supraglacial and englacial hydrology are highly complex, but crudely represented in ice sheet models. For example, the presence of surface meltwater does not necessarily lead to immediate ice shelf collapse (Bell et al., 2017; Kingslake et al., 2017), but for the ice shelves which have collapsed, surface meltwater was a precursor (Scambos et al., 2004; Banwell et al., 2013). Edwards et al. (2018) used statistical methods to excluding hydrofracturing and cliff instability from DeConto and Pollard (2016) results, and found that without these processes their model performs in line with Ritz et al. (2015).

Another fundamental limitation of all these studies is the lack of explicit interaction between the retreating ice sheet and the surrounding ocean (Asay-Davis et al., 2016; Seroussi et al., 2017). Massive freshwater inputs like those simulated by DeConto and Pollard (2016) during peak retreat in their RCP8.5 ensembles (> 1.5 Sv) would significantly alter sea ice, water column stratification, and ocean circulation surrounding the ice sheet, with plausible, albeit untested impacts on the amount of warm water penetrating ice shelf cavities to drive basal melt. Impacts would also be felt by the overlying atmosphere, with feedbacks on the trajectory of Antarctic surface climate. Accounting for ocean-ice interactions at a continental scale continues to be a major modelling challenge, requiring higher resolution ocean models than presently used. In light of uncertainties in both atmospheric and ocean forcing, and glacial hydrology, there is *low confidence* in the projected timing of widespread ice shelf collapse, but significant collapses in the second half of 21st century cannot be excluded.

The MICI mechanism was not considered in quantitative ice loss estimates by AR5, and it adds substantially to the dynamical component of ice loss, previously assumed to be limited to deformation, basal sliding, and calving. We stress that hydrofracturing and ice-cliff processes have only been included in one continental ice sheet model. In reality, mechanical ice failure is controlled by many interacting processes, including the stress regime at the ice front, water depth, ice thickness, flow speed, conditions at the bed of the ice, preexisting crevasses, lateral shear, undercutting of the calving face, and tides among others. The presence of mélange (a mix of previously calved, broken icebergs and sea ice) could also provide some buttressing support to a retreating cliff face providing a negative feedback (Amundson et al., 2010). Including the backstress provided by mélange is shown have little impact on the rate of large scale ice sheet retreat (Pollard et al., 2018), but these processes have not been directly accounted for in enough models to draw any conclusions at this time. DeConto and Pollard (2016) limited the rate of MICI retreat within the range demonstrated by a few thick, marine-terminating glaciers on Greenland (Howat et al., 2008; Joughin et al., 2008), however due to the general lack of observations and mechanistic, process-based modeling to date, the pace of sustained ice loss this process can produce at the special scale of an Antarctic outlet glaciers like Thwaites remains fundamentally unknown.

The continental-scale modeling studies of Ritz et al. (2015), Golledge et al. (2015), and DeConto and Pollard (2016) vary considerably in their approaches and their projections of Antarctica's future contribution to GMSL. However, they all represent a considerable departure from AR5, demonstrating 0.3 m or more of sea level rise from Antarctic by 2100 is possible for RCP8.5. For comparison, AR5 (Church et al., 2013), reported RCP4.5 and RCP8.5 median values (and *likely* ranges) of 0.10 m (0.03–0.19) and 0.12 m (0.03–0.20) in their assessment that considered rapid ice dynamics independent of forcing scenario, while adding the following: 'Based on current understanding, only the collapse of marine-based sectors of the Antarctic ice sheet, if initiated, could cause global mean sea level to rise substantially above the *likely* range during the

21st century. This potential additional contribution cannot be precisely quantified but there is *medium confidence* that it would not exceed several tenths of a meter of sea level rise during the 21st century' (Church et al., 2013). Given the publications after AR5 we reassess Antarctica's contribution of sea level. Firstly, we now conclude that strong divergence in the forcing among the different scenarios means that projecting the dynamic contribution independent of the RCP scenario, as done by Church et al. (2013) due to a lack of literature on the topic, is no longer justified. All recent studies indicate a very limited contribution of Antarctica to sea level for the RCP2.6 scenario (*medium confidence*). For the higher RCP8.5 scenario, the difference between the different studies is considerable, leading to a much larger uncertainty.

The assessment of Antarctica's future contribution to GMSL is based on, first, averaging the two alternative subgrid parameterization scenarios of Golledge et al. (2015) for the A1B scenario. The results are in good agreement. This provides support for giving the Golledge et al. (2015) considerable weight in assessing the Antarctic contribution for RCP8.5. Secondly, we compare the result from Golledge et al. (2015) for RCP8.5 (0.1–0.39 m in 2100) to the low Pliocene high-stand calibrated results by DeConto and Pollard (2016 0.64 m in 2100). The latter study includes processes in addition to MISI (hydrofracturing and marine ice cliff instability) that make an additional but highly uncertain contribution to sea level rise. Consequently, we give the latter study less weight than Golledge et al. (2015) and assess the *likely* Antarctic contribution to GMSL as closer to the mean of the Golledge et al. (2015) result, 0.3 ± 0.1 m (one sigma uncertainty) for the period 2081–2100. [PLACEHOLDER FOR SECOND ORDER DRAFT: several studies are underway to provide further guidance on this number.]

Some studies based on expert elicitation point to a non-Gaussian distribution particular after 2100, but process based and observational studies like Ritz et al. (2015) and DeConto and Pollard (2016) do not provide compelling evidence for this. Hence, we assume a Gaussian distribution for this century. The estimated end-of-the-century contribution for Antarctica includes SMB and ice-sheet dynamics. Results of the model simulations after 2100 are considered to be too uncertain to include in the *likely* range, but are discussed in Section 4.2.3.6.

Table 4.2: An overview of different studies to estimate the MISI contribution to sea level rise. [PLACEHOLDER FOR SECOND ORDER DRAFT: Tables and Figures below to be revised to account for additional literature, e.g., Edwards et al., 2018; DeConto et al., 2018; Levermann et al., 2018; possibly others]. The values for DeConto and Pollard (2016) represent ensembles calibrated to their low range of Pliocene sea level targets (5–15 m) with the use of an ocean temperature bias correction in the Amundsen Sea.

	Ritz et al. (2015)	Golledge et al. (2015)	DeConto and Pollard (2016)
	RCP2.6/RCP4.5/ A1B/RCP8.5	*RCP2.6/RCP4.5/ A1B/RCP8.5*	*RCP2.6/RCP4.5/ A1B/RCP8.5*
GMSL[*–1] 2050 (m)	-/-/0.03/-	0.04/0.05/-/0.07	0.02/0.03/-/0.04
GMSL 2100 (m)	-/-/0.12/-	0.06/0.10/-/0.23	0.14/0.41/-/0.79
GMSL 2200 (m)	-/-/0.41/-	0.13/0.37/-/1.20	0.35/1.67/-/5.39
Uncertainties	Quantiles	High-Low	Ensemble selections
Tuning targets	Present-day rates from observations	None	Last Interglacial and Pliocene
Grounding Line	Conditional on bed slope and Schoof flux	Sub-grid parameterization	Pollard and DeConto (2012)
Dynamics	Several basal friction laws	Hybrid, 10-20 km grid Till friction angle	Hybrid, 10 km grid
Hydrofracturing	No	No	Yes
Marine Cliff Instability	No	No	Yes
Initialization	Observed rates	Focus on long time scales	1950
SMB	parameterized	PDD scheme	Regional Climate Model
BMB	parameterized	Slab Ocean GCM	NCAR CCSM4
Driving mechanism for retreat	Observations, statistics	Ocean (2/3)	Atmospheric forcing

4.2.3.2 Global Projections of Sea Level Rise

The estimated values for the MISI contribution in 2081–2100 imply that the contribution from Antarctica is considered to be somewhat larger than the thermal expansion from CMIP5 simulations (Church et al., 2013) and with a larger uncertainty as well. There is limited evidence for major changes since AR5 in the other (Glaciers, Greenland, Thermal expansion and land water storage) components to sea level rise, partly caused by a lack of new CMIP simulations. Hence, we have constructed new projections by replacing the AR5 estimate for Antarctica by a new assessment as outlined in the previous paragraph and maintaining similar contributions for the other components. Results are shown in Table 4.3.

Table 4.3: Median values and *likely* ranges for projections of global mean sea level (GMSL) in meters in 2081–2100 relative to 1986–2005 for three scenarios. In addition, rates for 2046–2055 are mentioned as well as the GMSL in 2100 and the rate of GMSL in 2100. Values between square brackets reflect the *likely* range. [PLACEHOLDER FOR SECOND ORDER DRAFT: results may change in case more studies can be used to estimate MISI]

(meters)	RCP2.6	RCP4.5	RCP8.5	Comments
Thermal expansion	0.14 (0.10–0 .18)	0.19 (0.14–0.23)	0.27 (0.21–0.33)	AR5
Glaciers	0.10 (0.04–0.16)	0.12 (0.06–0.18)	0.16 (0.09–0.23)	AR5
Greenland SMB	0.03 (0.01–0.07)	0.04 (0.02–0.09)	0.07 (0.03–0.17)	AR5
Greenland DYN	0.04 (0.01–0.06)	0.04 (0.01–0.06)	0.05 (0.02–0.07)	AR5
LWS	0.04 (0.01–0.06)	0.04 (0.01–0.06)	0.04 (0.01–0.06)	AR5
Total - Antarctica AR5; 2081–2100	0.34 (0.26–0.44)	0.42 (0.33–0.53)	0.59 (0.47–0.73)	SROCC implicit in AR5
Total AR5; 2046–2065	0.24 (0.18–0.30)	0.26 (0.20–0.32)	0.29 (0.23–0.36)	SROCC implicit in AR5
Total AR5 - Antarctica AR5; 2046–2065	0.22 (0.17–0.27)	0.23 (0.19–0.29)	0.27 (0.22–0.33)	SROCC implicit in AR5
Antarctica 2046–2065	0.01 (0.01–0.01)	0.02 (0.01–0.03)	0.05 (0.02–0.09)	SROCC
Antarctica 2081–2100	0.06 (0.04–0.08)	0.12 (0.06–0.18)	0.30 (0.1–0.5)	SROCC
GMSL 2046–2065	0.26 (0.20–0.31)	0.27 (0.22–0.33)	0.31 (0.26–0.37)	SROCC
GMSL 2081–2100	0.40 (0.31–0.51)	0.54 (0.43–0.66)	0.89 (0.66–1.13)	SROCC
GMSL 2200	Outside *likely* range	Outside *likely* range	Outside *likely* range	
GMSL in 2100	0.44 (0.33–0.56)	0.59 (0.47–0.73)	1.06 (0.82–1.33)	SROCC
Rate (mm yr^{-1})	5	8	18	SROCC

Results as presented in Table 4.3 are used to calculate the regional RSL projections are used in 4.2.3.4 to calculate extreme sea level projections. Time series for the different RCP scenarios are shown in Figure 4.7 clearly indicating a divergence in both magnitude and uncertainty between this report and the AR5 projections (Church et al., 2013) for the higher RCP scenarios.

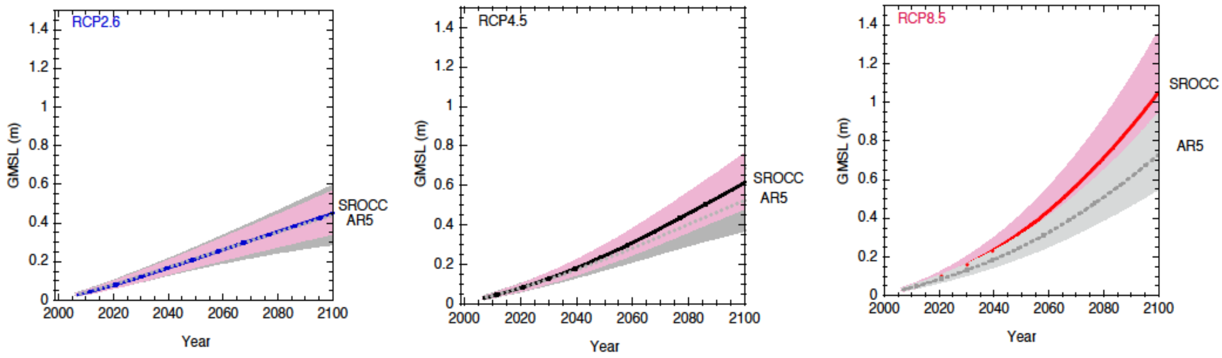


Figure 4.7: Time series of GMSL for RCP2.6, RCP4.5 and RCP8.5 as used in this report as well as for reference the AR5 results (Church et al., 2013). Results are based on AR5 results for all components except the Antarctic contribution. Results for the Antarctic contribution in 2081–2100 are provided in Table 4.3. All components are treated independently and the shaded area is the 17–83% confidence interval, which is considered to be the *likely* range.

Including the updated results in terms of magnitude and uncertainty for the Antarctic component also changes the regional patterns in sea level projections. Results of the regional patterns as shown in Figure 4.8 show an increased sea level rise w.r.t. results presented in AR5 nearly everywhere for RCP8.5 because of the increased Antarctic contribution. Differences are largest along the diagonal from the Indian Ocean to the North-Atlantic Ocean as a result of gravitational and rotational effects.

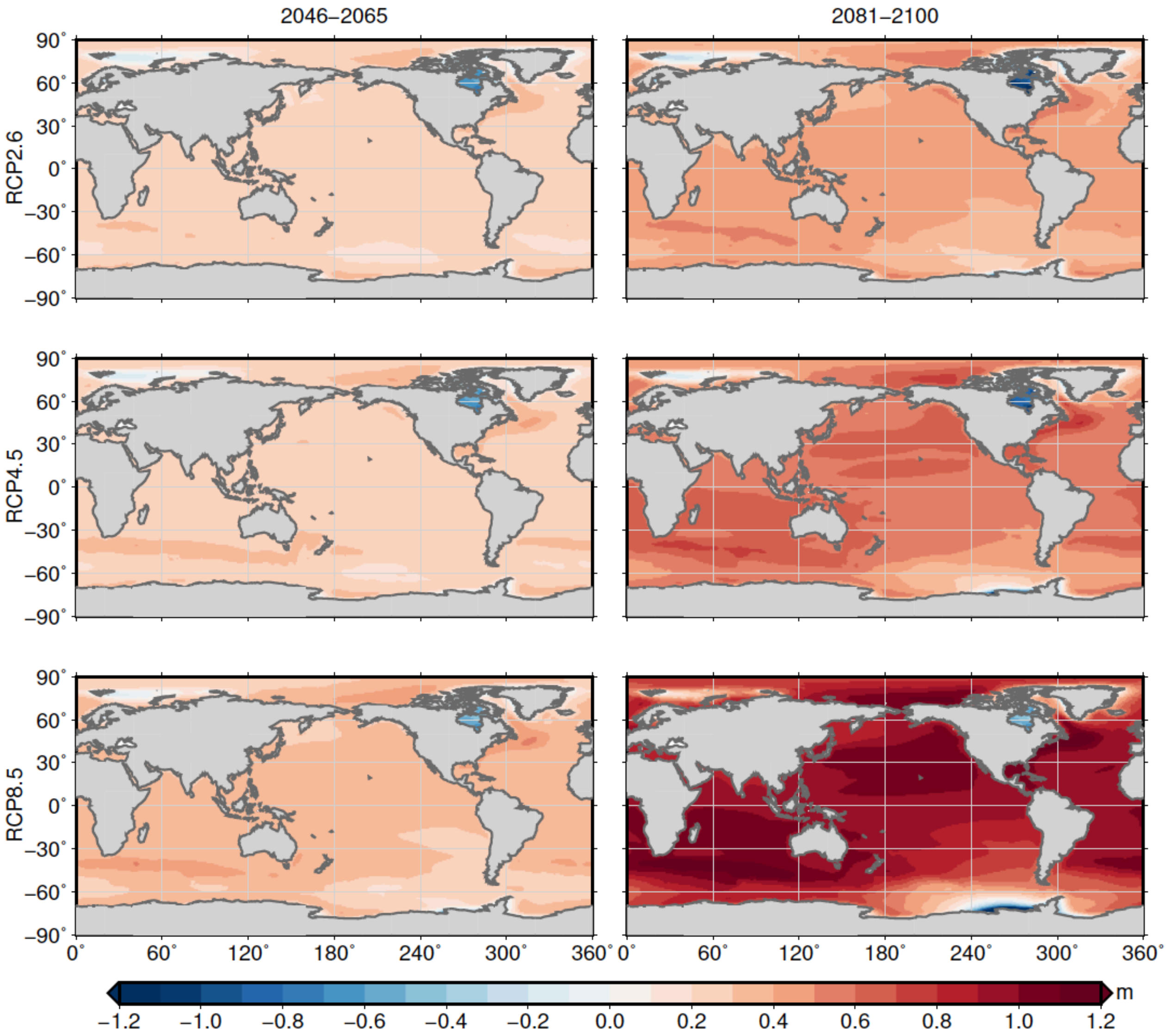


Figure 4.8: Regional relative sea level change for RCP2.6, RCP4.5 and RCP8.5 as used in this report for ESL calculations. Results are based on AR5 results except for the Antarctic contribution. The left column is for the time slice 2046–2065 and the right column for 2081–2100 the magnitude of the sea level rise in meter, the right column the standard error. Results are presented for the difference between 2081–2100 and 1986–2005. The supplementary information shows the results for 2046–2055 and the details of the calculations.

4.2.3.3 *Probabilistic Sea Level Projections*

Since AR5, several studies have produced sea level rise projections in coherent frameworks that link together global-mean and local relative sea level rise projections. The approaches are generally similar to those adopted by AR5 for its global-mean sea level projections: a bottom-up accounting of different contributing processes (e.g., land-ice mass loss, thermal expansion, dynamic sea level), but many also are 'probabilistic', in that they attempt to describe more comprehensive probability distributions of sea level change than the '*likely*' ranges presented by Church et al. (2013). An example is the study by Le Bars et al. (2017) who expand the projection by Church et al. (2013) in a probabilistic way with the Antarctic projections by DeConto and Pollard (2016) to obtain a full probability density function for sea level rise. These estimates are necessary for incorporation in a quantitative risk management framework (see Section 4.3.3). An even more general approach has been taken by Le Cozannet et al. (2017) who frame a probabilistic framework of sea level rise including all existing probabilistic estimates.

This section first briefly reviews key sources of information for probabilistic projections (Section 4.2.3.3.1), with a focus on new results since AR5, then summarizes the different global and regional projections (Section 4.2.3.3.2). Eventually, we distinguish bottom-up projections which explicitly describe the different component to sea level rise (Section 4.2.3.3.3) and semi-empirical projections (Section 4.2.3.3.4).

4.2.3.3.1 Components of probabilistic GMSL projections

Thermal expansion: Global mean thermal expansion projections rely on AOGCM projections (Kopp et al., 2014; Slangen et al., 2014a; Jackson and Jevrejeva, 2016) or simple climate model projections (Perrette et al., 2013; Bakker et al., 2017b; Nauels et al., 2017b), and are substantively unchanged since AR5. For those studies relying on the CMIP5 AOGCM ensemble, interpretations of the model output differ mainly with regard to how the range is understood, e.g., Kopp et al. (2014), interprets the 5th–95th percentile of CMIP5 values as a *likely* range of thermal expansion. The differences among the studies yield discrepancies smaller than 10 cm. For example, Slangen et al. (2014b) project a 1σ range under RCP8.5 of 20–36 cm in 2081–2100 vs. 1986–2005, while Kopp et al. (2014) project a *likely* range of 28–46 cm in 2081–2099 vs 1991–2009. Little et al. (2015b) note that the 'crossover time', at which scenario-driven uncertainty in global mean thermal expansion becomes larger than internal variability, occurs in about 2035. This is partially explained by Melet and Meyssignac (2015) who show that 30% of the spread in the projection is caused by differences in the forcing, 35% differences in climate sensitivity among the models and 35% is due to the spread in ocean heat uptake among the models.

Glaciers: Projections of glacier and ice cap mass change rely either on models of glacier surface mass balance and geometry, forced by temperature and precipitation fields (Slangen and Van de Wal, 2011; Marzeion et al., 2012; Hirabayashi et al., 2013; Radić et al., 2014; Huss and Hock, 2015), or simple scaling relationships with global mean temperature (Perrette et al., 2013; Bakker et al., 2017b; Nauels et al., 2017b). Glacier mass change projections published since AR5, based on newly developed glacier models, confirm the overall assessment of AR5 (see also Section 4.2.3.2).

Land water storage: Projections of the GMSL rise contributions due to dam impoundment and groundwater withdrawal are generally either calibrated to hydrological models (e.g., Wada et al., 2012) or neglected. Recent coupled climate-hydrological modelling suggests that a significant minority of pumped groundwater remains on land, which may reduce total GMSL rise relative to studies assuming full drainage to the ocean (Wada et al., 2016). However, there are no substantive updates to projections of the future land-water storage contribution to GMSL rise since AR5.

Ice sheets: Existing GMSL projections rely upon some combination of (1) past expert assessments by the IPCC based on physical models of varying degree of complexity (Meehl et al., 2007; Church et al., 2013) and other forums (Katsman et al., 2011), or alternatively (2) structured expert elicitation. Approach (1) is based on CMIP5 calculations of which the uncertainty range was assessed to be larger than the direct uncertainty range from the models yielding that the 5%–95% range was interpreted as the *likely* range from 17%–83%. Approach (2) adopted a more formal expert elicitation protocol (Cooke, 1991) rather than physical based models. Those results yielded a significant higher contribution of the ice sheets to sea level rise, but results were criticized for their adopted methodology of post-processing the expert data (de Vries and van de Wal, 2015; de Vries and van de Wal, 2016). Alternatively, Horton et al. (2014) used a simpler elicitation protocol focusing on the total sea level rise rather than the ice sheet contribution with results more in line with earlier IPCC assessments. Beside the total contribution of ice sheets several studies address the contribution of either Greenland or Antarctica (see Section 4.2.3.1.1 and 4.2.3.1.2). Critical for GMSL projections is the *low confidence* in the dynamic contribution of the Antarctic ice sheet beyond 2050 as discussed in Section 4.3.2.1.2.

4.2.3.3.2 From probabilistic GMSL projections to regional relative sea level change

Differences between GMSL and relative sea level change are driven by three main factors: (1) dynamic sea level (DSL), for instance, the thermal expansion component and the circulation driven changes, (2) GIA effects often separated into instantaneous gravitational and rotational effects caused by redistribution of mass within cryosphere and hydrosphere, leading to fingerprint patterns, and 3) long term processes that lead to vertical land motion. Finally, the inverse barometer effect caused by changes in the atmospheric pressure, sometimes neglected in projections, can also make a small contribution, particular on shorter time scales. For the 21st century as a whole estimates are smaller than 5 cm at local scales (Carson et al., 2016).

Dynamic sea level (DSL): Projections of dynamic sea level change are necessarily derived through interpretations of AOGCM projections. As with thermal expansion projections, interpretations of the CMIP5 ensemble differ with regard to how the model range is understood and the manner of drift correction, if any (Jackson and Jevrejeva, 2016). However, relative to tide-gauge observations, AOGCMs tend to overestimate the memory in dynamic sea level; thus, they may underestimate the emergence of the externally forced signal of DSL change above scenario uncertainty (Becker et al., 2016).

Gravitational-rotational effects: All projections of relative sea level change include fingerprints for cryospheric changes, though they differ in the details with which these are represented. Some studies also include a fingerprint for land-water storage change (Slangen et al., 2014b). Recent work indicates that, for some regions with low mantle viscosity, fingerprints cannot be treated as fixed on multi-century timescales (Hay et al., 2017). This effect has not yet been incorporated into comprehensive RSL projections, but is probably only of relevance near ice sheets. We have *high confidence* in the patterns caused by gravitational and rotational effect, as in AR5.

Long term solid Earth processes: These processes can be an important driver of relative sea level change, particularly in the near- to intermediate-field of the large ice-sheets of the Last Glacial Maximum (e.g., North America and northern Europe). Studies differ as to whether this process is incorporated by physical modeling (e.g., Slangen et al., 2014b) or by estimation of a long-term trend from tide-gauge data (e.g., Kopp et al. (2014), which is then spatially extrapolated). In the former case, projections may exclude other important local factors contributing to vertical land motion (e.g., tectonic uplift/subsidence and groundwater/hydrocarbon withdrawal); in the latter, projections may assume that these other processes proceed at a steady rate and thus do not allow for management changes that affect groundwater extraction.

4.2.3.3.3 Probabilistic bottom-up projections

A wide range of probabilistic sea level projections exist, ranging from simple scaling relations to partly process-based components combined with scaling relations. Table 4.4 illustrates the supplementary level of many of the studies. Many rely for an important part of their components on CMIP simulations and the largest difference can be found on the treatment of the ice dynamics, particularly for Antarctica which are usually not CMIP5 based. Instead, each derives from one of several estimates of the Antarctic contribution. These results are extremely useful for the purposes of elucidating sensitivities and bounds. In SROCC, we rely on the Antarctic component from 4.2.3.2 for calculating the *likely* range of RSL because it is based on assessment of multiple studies, excluding MICI as this process is assessed to be deeply uncertain. Comparing the probabilistic projections is difficult because of the subtle differences between their assumptions. Nevertheless, values of sea level rise are presented in Table 4.5. Typically, values range much more for 2100 then for 2050.

Table 4.4: Sources of Information Underlying Bottom-up Projections of Sea level Rise Projections. (TE= Thermal Expansion, Glaciers, LWS=Land water storage changes, DSL=Dynamic Sea Level, GIA+VLM=Glacial Isostatic Adjustment and Vertical Land Motion).

Study	TE	Glaciers	LWS	Ice Sheets	DSL	Fingerprint	GIA + VLM
Perrette et al. (2013)	CMIP5	Global surface mass balance	Not included	Greenland's surface mass	CMIP5	Bamber et al. (2009)	Not included

		sensitivity and exponent from AR4; total glacier volume from Radić and Hock (2010)		balance from AR4; semi-empirical model using historical observations.			
Grinsted et al. (2015)	CMIP5	AR5 projections	Wada et al. (2012)	AR5 projections; Expert elicitation from Bamber and Aspinall (2013)	CMIP5	Bamber et al. (2009)	GIA projections from Hill et al. (2010) using observations
Slangen et al. (2014b)	CMIP5	CMIP5; glacier area inventory Radić and Hock (2010) in a glacier mass loss model	Wada et al. (2012)	SMB Meehl et al. (2007), ice dynamics Meehl et al. (2007) and Katsman et al. (2011)	CMIP5	Mitrovica et al. (2001)	GIA resulting of ice sheet melt from glacier mass loss model
Kopp et al. (2014)	CMIP5	CMIP5; Marzeion et al. (2012)	Chambers et al. (2017); Konikow (2011)	AR5 projections; Expert elicitation from Bamber and Aspinall (2013)	CMIP5	Mitrovica et al. (2011)	GIA, tectonics, and subsidence from Kopp et al. (2013)
Jackson and Jevrejeva (2016)	CMIP5	Marzeion et al. (2012)	Wada et al. (2012)	AR5 projections; Expert elicitation from Bamber and Aspinall (2013)	CMIP5	Bamber et al. (2009)	GIA resulting of ice sheet melt from glacier mass loss model Peltier et al. (2015)
Horton et al. (In Press)	CMIP5	CMIP5; Marzeion et al. (2012)	Wada et al. (2012); Konikow (2011)	AR5 projections; Expert elicitation from Bamber and Aspinall (2013); DeConto and Pollard (2016)	CMIP5	Mitrovica et al. (2011)	GIA, tectonics, and subsidence from Gaussian-process model
De Winter et al. (2017)	CMIP5	CMIP5; glacier area inventory Radić and Hock (2010) in a glacier mass loss model	Wada et al. (2012)	AR5 projections; Expert elicitation de Vries and van de Wal (2015); Ritz et al. (2015)	CMIP5	Mitrovica et al. (2001)	GIA resulting of ice sheet melt from glacier mass loss model

Table 4.5: Median and *likely* GMSL rise projections (m). Values between brackets are *likely* range, if no values are given the *likely* range is not available.

	2050			2100		
	RCP2.6	RCP4.5	RCP8.5	RCP2.6	RCP4.5	RCP8.5
Perrette et al. (2013)	---	0.28 (0.23–0.32)	0.28 (0.23–0.34)	---	0.86 (0.66–1.11)	1.06 (0.78–1.43)
Grinsted et al. (2015)	---	---	---	---	---	0.8 (0.58–1.20)
Slangen et al. (2014b)	---	---	---	---	0.57 (0.37–0.77)	0.74 (0.45–1.04)
Kopp et al. (2014)	0.25 (0.21–0.29)	0.26 (0.21–0.31)	0.29 (0.24–0.34)	0.50 (0.37–0.65)	0.59 (0.45–0.77)	0.79 (0.62–1.00)
Jackson and Jevrejeva (2016)	---	---	---	---	0.52 (0.34–0.69)	0.72 (0.52–0.94)
Horton et al. (In Press)	0.23 (0.16–0.33)	0.26 (0.18–0.36)	0/31 (0.22–0.40)	0.56 (0.37–0.78)	0.91 (0.66–1.25)	1.46 (1.09–2.09)
Nauels et al. (2017b)	0.22 (0.17–0.27)	0.23 (0.19–0.28)	0.25 (0.20–0.30)	0.43 (0.34–0.54)	0.52 (0.43–0.63)	0.71 (0.58–0.87)
Nauels et al. (2017a)	0.20 (0.14–0.29)	---	0.25 (0.18–0.33)	0.49 (0.33–0.71)	0.67 (0.43–0.99)	0.88 (0.59–1.27)
Bakker et al. (2017b)	0.18	0.21	0.23	0.51	0.68	1.11
Wong et al. (2017)	0.26	0.28	0.30	0.55	0.77	1.50
Jevrejeva et al. (2012)	---	0.29	0.33	---	0.67	1.00
Schaeffer et al. (2012)	---	---	---	---	0.90	1.02
Mengel et al. (2016)	0.17	0.17	0.19	0.38	0.51	0.81
De Winter et al. (2017)	---	---	---	---	---	0.68/0.86

4.2.3.3.4 Semi-empirical projections

Semi-empirical models provide an alternative approach for process-based models aiming to close the budget between observed sea level rise and the sum of the different components contributing to sea-rise, and secondly related to this, try to quantify the contribution of ice sheet dynamics to sea level rise. In general, semi-empirical models use statistical correlations, motivated by a mechanistic understanding, from time series analysis of observations to generate projections. They implicitly assume that processes driving the observations are operating similarly in the past as they do in future including similar feedback mechanisms. With the advances in closing the sea level budget and in our process understanding of the dynamics of ice sheet processes the values of semi-empirical models gradually decays particularly as it is now realized that the dynamical changes driving changes in Antarctica are poorly or not captured in the recent observations. MISI may have a very different character in the near future than in the recent past and hydrofracturing is impossible to quantify on observational records only. Moreover, their results (e.g., Kopp et al., 2016; Mengel et al., 2016; Mengel et al., 2018) are in general agreement with Church et al. (2013). Only if they include specific estimates of the dynamic contribution of Antarctica which strongly deviate from the values adopted by Church et al. (2013) like the combined hydrofracturing and ice cliff instability mechanism as presented by DeConto and Pollard (2016), total sea level projections deviate as well (e.g., Nauels et al.2017).

4.2.3.4 Extreme Sea Level Projections

Extreme sea levels (ESL) events, also known as storm tides, are water level heights that consist of contributions from mean sea level, storm surges, tides, and waves. Even a small increase in mean sea level can significantly augment the frequency and intensity of flooding. This is because SLR elevates the platform for storm surges, tides, and waves, and because there is a log-linear relationship between a flood's height and its occurrence interval. For example, tidal changes are non-linearly related to mean high water so that, for fixed coast lines in 10% of the coastal cities, variability in projected mean high water is larger than the variability in SLR itself by 10% (Pickering et al., 2017). Changes are most pronounced in shelf seas. Over 300 million people reside in areas that are exposed to ESL, experiencing tens of billions of dollars in damages per year (Wahl et al., 2017). Roughly 1.3% of the global population is exposed to a 1/100-year flood (Muis et al., 2016). This exposure to ESL and its damage could increase significantly with SLR, potentially amounting to 10% of the global gross domestic product by the end of the century without adaptation (Wahl et al., 2017).

Frequencies of ESL events can be estimated with hydrodynamic or statistical models. Hydrodynamic models simulate a series of ESL over time, which can then be fitted by extreme value distributions to estimate the frequency and intensity of ESL (e.g., the height of the 1/100-year flood). Statistical models fit tide gauge observations to extreme value distributions to directly estimate storm tide distributions. Both of these modelling approaches can account for projections of SLR. The statistical models implicitly assume that the extreme values distribution is not changing over time. An advantage of the use of hydrodynamic models is that they can quantify interactions between the different components of ESL. Hydrodynamical models can be executed over the entire ocean with flexible grids at a high resolution (up to 1/20° or ~5 km) where necessary, appropriate for local estimates (Kernkamp et al., 2011). Input for those models are wind speed and direction, and atmospheric pressure. Results of those models show that the Root Mean Squared Error RMSE between modelled and observed sea level is less than 0.2 m for 80% of a data set of 472 stations covering the global coastline (Muis et al., 2016) at 10 minute temporal resolution over a reference period from 1980–2011. Although the model poorly represents tropical storms, this accuracy implies that this type of model may adequately describe the variability in ESL. Hydrodynamical models often contain a tidal model component to improve projections of ESL. Several tidal models exist, which perform well for present-day conditions (Stammer et al., 2014). [PLACEHOLDER FOR SECOND ORDER DRAFT: new work in progress on hydrodynamical models may be captured].

Statistical models have shown that the estimation of ESL is highly sensitive to the characterization of SLR and flood frequency distributions (Buchanan et al., 2017). This is confirmed by Wahl et al. (2017) who estimate that the 5–95 percentile uncertainty range, attained through the application of different statistical extreme value methods and record lengths, of the current 1/100-year flood event is on average 40 cm, whereas the corresponding range in projected GMSL of AR5 under RCP8.5 is 37 cm. For ESL events with a shorter return period, differences will be even larger. Capturing changes in the ESL return periods in the future is even more complicated because both the changing variability over time and the uncertainty in the mean projection have to be combined. A statistical framework to combine RSL and ESL, based on historical tide gauge data was applied to the U.S. coastlines (Buchanan et al., 2016). Hunter (2012) and the AR5 (Church et al., 2013) projected changes in flood frequency worldwide; however, these analyses used the Gumbel distribution for high water return periods, which implies that the frequency of all ESLs (e.g., whether the 1/10-year or 1/500-year) will change by the same magnitude for a given sea level rise, and thus can underestimate or overestimate ESL (Buchanan et al., 2017). Hence, the amplification factors of future storm return frequency in AR5 WGI Figure 13.25 may underestimate flood hazards in some areas, while overestimating them in others. By using the Gumbel distribution, Muis et al. (2016) may also inadequately estimate flood frequencies.

4.2.3.4.1 RSL and ESL projections based on tide gauge records

Here we present results in ESL based on the global projections as presented in 4.3.2.1 for the tide gauges in the GESLA2 database (Woodworth et al., 2016). Return periods are calculated as a combination of regional relative sea level changes and a characterization of the variability in sea level as derived from the GESLA2 data set which contains tide gauges from all over the world. By doing so it is assumed that the variability as characterized by the tide gauges is not changing over time. To accommodate the non-exponential character of the relation between return height and return period in the tide gauges the relation is characterized by a Generalized Pareto Distribution (GPD) fit on declustered (72 hours between the peaks) tide gauge observations, with a threshold value of 99.7%, basically following the approach by Arns et al. (2013). Uncertainties are estimated by a Monte Carlo approach (see Supplementary Material for details). To estimate future return heights and periods the fits from the tide gauges are combined with the RSL as presented in 4.3.2.1 for RCP2.6, RCP4.5 and RCP8.5 for two periods: 2046–2065 and 2081–2100. Results are shown for 12 selected tide gauges in Figure 4.9. Depending on the curvature of the relation between return height and return period future extreme level conditions are determined by the regional relative sea level increase or by the current variability in sea level arising from the tide gauge record, see inset Figure 4.9. If the difference between mean sea level and typical extremes is large (e.g., Cuxhaven) the regional relative sea level rise is less important and vice versa (e.g., Kerguelen Island). For Balboa, Sydney and Kerguelen Island, the projected relative-sea level rise is so large that the return heights are above the return heights as measured by the tide gauges (black crosses in Figure 4.9). A discrepancy exists between locations that experience rare and very high ESLs (such as Papeete and New York), and those that have a statistical upper bound to ESLs (such as Kerguelen Island and Sydney). The rare, very high historical ESLs are often found in regions where cyclones and hurricanes occur. At these locations, the PDRP and AF are largest for relatively common events, but these values are limited for rare events. The statistical upper bound to ESLs are often found in regions where tidal variability is large with respect to storm surges. In these locations, sea level rise will particularly increase the occurrence frequency of rare events.

More importantly, most of the locations show that a return height with a historical return period of 1/100 are projected to occur more than once a year during future conditions. This is expressed in Figure 4.10 for three climate scenarios, mid-way the 21st century and at the end of the 21st century. From the curves of return height versus return period as shown in Figure 4.9, we can derive several quantities, which are relevant for decision making, being the PDRP_1 denoting the present-day return period that corresponds to the return height associated with a return period of 1 after a change in relative sea level, see inset in Figure 4.9; the SSLR_100 being the mean sea level change scaled by the present-day 100-yr return height. It is used in Cross Chapter Box 4; and the AF_100 is the amplification factor during changed conditions for events, which have during present-day conditions a return period of 100 years. Both PDRP and AF depend on the curvature of the relation between return height and return period and change as a function of return period. Results for each tide gauge station are provided in the supplementary information. Figure 4.10 shows the very large changes in extreme sea level over time as a function of three different RCP scenarios. The figure shows that, in many locations event which are currently have an estimated return period of a hundred years or more are increasing to a once every year event by the end of the century, particularly for the higher RCP8.5 scenario.

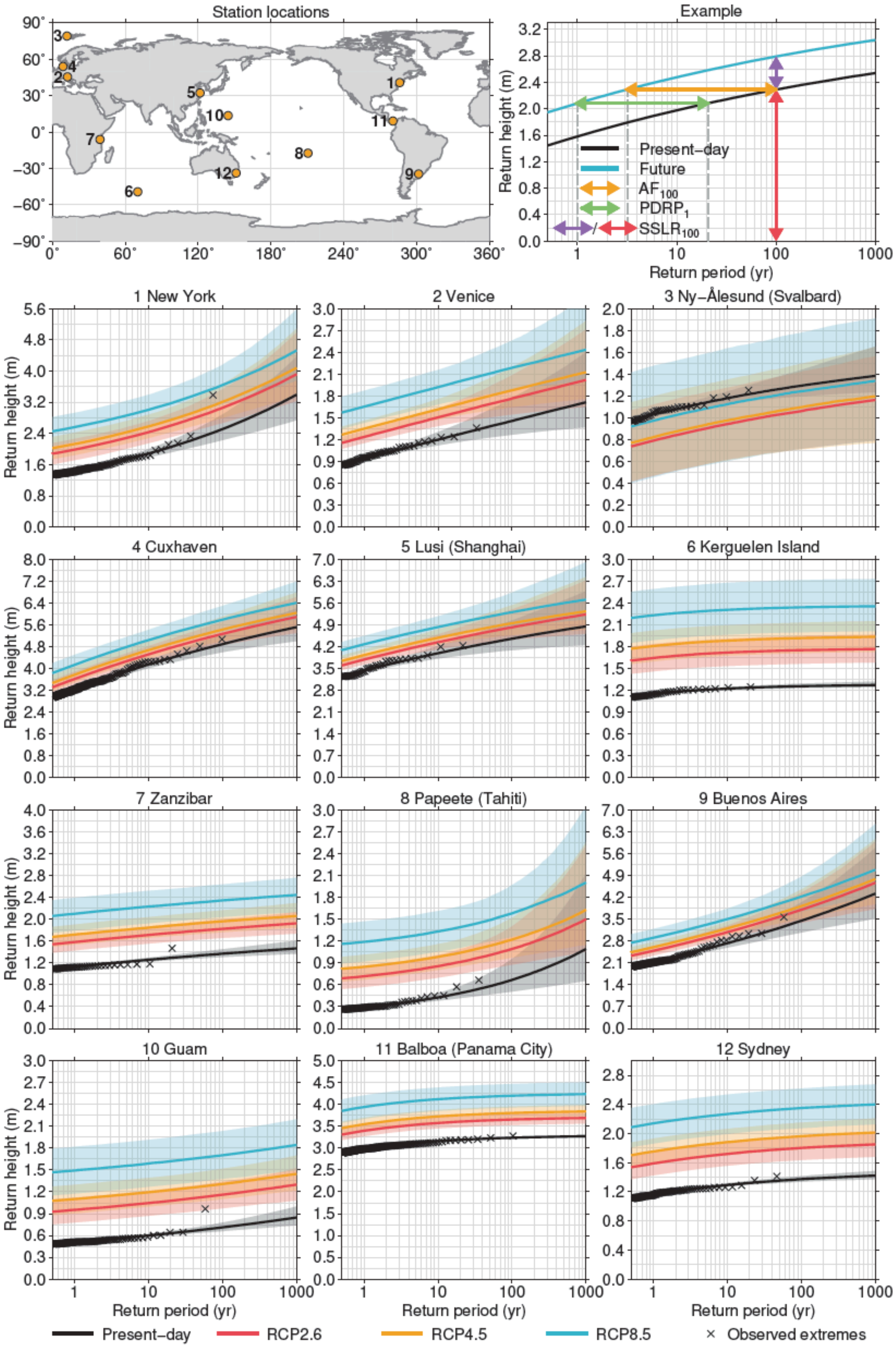

Station locations
90°
60°
30°
0°
–30°
–60°
–90°
0°
60°
120°
180°
240°
300°
360°
1
2
3
4
5
6
7
8
9
10
11
12
Example
Return height (m)
3.2
2.8
2.4
2.0
1.6
1.2
0.8
0.4
0.0
1
10
100
1000
Return period (yr)
Present–day
Future
AF_{100}
$PDRP_1$
/
$SSLR_{100}$
1 New York
2 Venice
3 Ny–Ålesund (Svalbard)
4 Cuxhaven
5 Lusi (Shanghai)
6 Kerguelen Island
7 Zanzibar
8 Papeete (Tahiti)
9 Buenos Aires
10 Guam
11 Balboa (Panama City)
12 Sydney
Return height (m)
Return period (yr)
1
10
100
1000
Present–day
RCP2.6
RCP4.5
RCP8.5
× Observed extremes

Figure 4.9: Return periods a set of characteristic tide gauge locations (see upper left for their location) for present-day condition (black lines) and 2081–2100 conditions for three different scenarios. The green line in the upper right panel is the AF100 being the amplification factor for 1/100 expressing the increase in occurrence frequency of events which has now a return period of 100 years. The PDRP_1 denotes the present-day return height that corresponds to the return height associated with a return period of 1 year after a change in relative sea level. The SSLR_100 is the relative sea level rise scaled by the present-day 100 year return period.

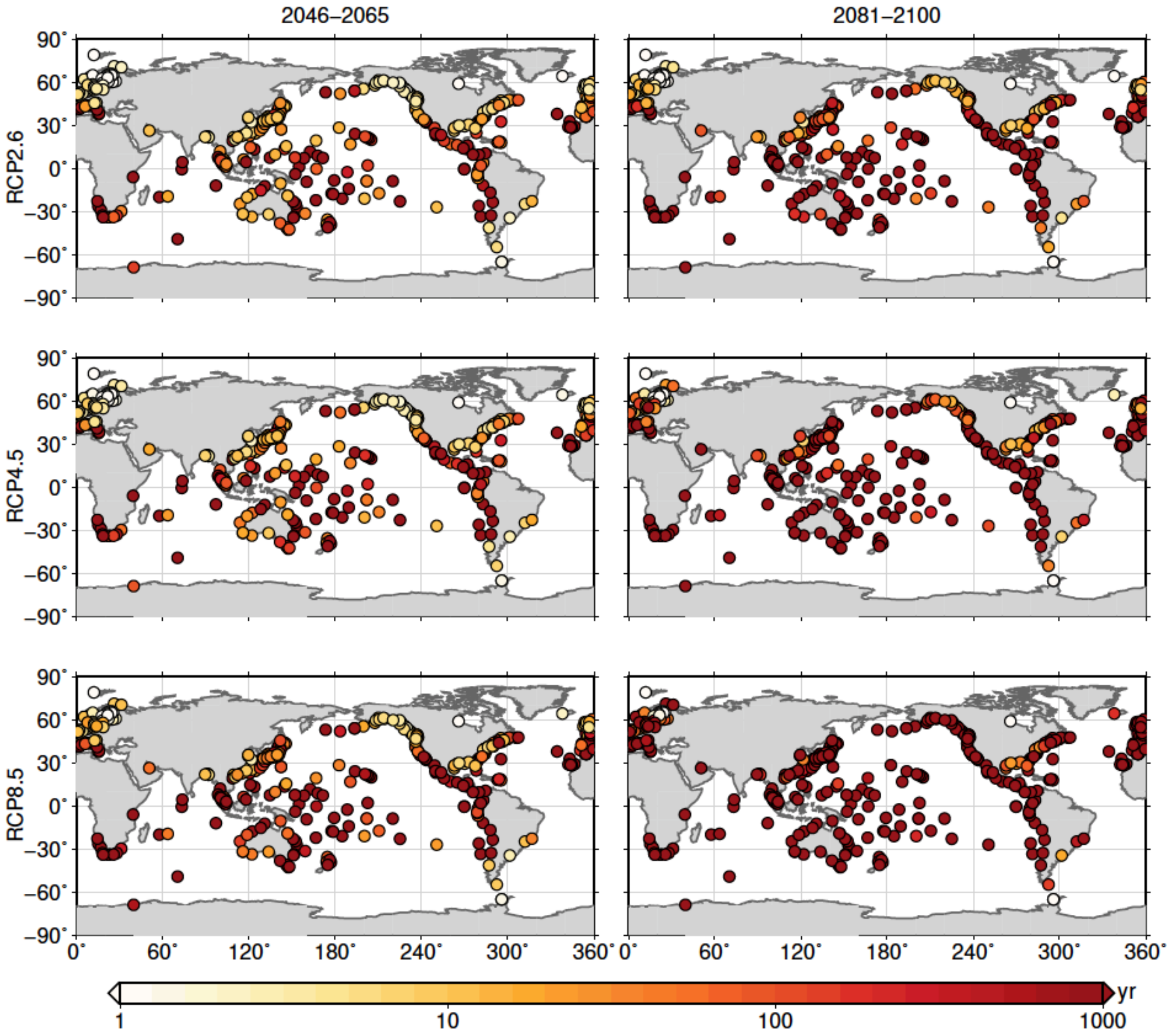


Figure 4.10: Present-day return period (PDRP_1) that correspond to a return height associated with a return period of 1 year after a change in relative sea level (i.e., is expected to occur annually). Results are shown for three RCP scenarios and two future time slices as median values. Results are shown for tide gauges in the GESLA2 database. The accompanying confidence interval can be found in the supplementary information.

In summary extreme sea level estimates as presented in this paragraph, clearly show that as a consequence of sea level rise, events which are currently rare (e.g., with an expected return period of 100 years), will occur yearly or more frequently at many locations for RCP8.5 by the end of the century (*high confidence*). For some locations, this change will occur already by mid-century for RCP8.5 and by 2100 for all emission scenarios. These locations are particularly located in low-latitude regions, away from the storm tracks. In these locations, historical sea level variability due to tides and storm surges is small compared to projected sea level rise. Therefore, even limited changes in mean sea level will have a noticeable effect on ESLs, and for some locations, even RCP2.6 will lead to the annual occurrence of historically rare events by mid-century.

4.2.3.4.2 Waves

Besides surges, flooding is also caused by change in waves which are usually not included in hydrodynamical models and also not in the calculations presented in Figure 4.10. Wang et al. (2014b) show that the annual mean significant wave height (wave height trough to crest of the highest third of the waves) as calculated based on 6 hourly data with 20 CMIP5 models forced with the RCP8.5 scenario, increases in tropical regions as a result of increased sea level pressure gradients and stronger surface winds. The impact of future sea level rise and coastal management strategies on the tides is also relevant. In addition Arns et al. (2017) show that an increase in sea level reduces the depth-limitation of waves, thereby resulting in waves with greater amplitude, period and higher run-up. In combination with changes in the tides design heights needs to be increased by 48%–56% in the German bight region relative to a design height based on sea level rise only. For the Southern North Sea region Weisse et al. (2012) argue that an increasing storm activity also increases hazards from extreme sea levels. Vousdoukas et al. (2017) quantify the extreme sea level including a wave model to be nearly 1 m under RCP8.5 in the North Sea region which is the highest in Europe. This is a 40% increase of the RSLR trends caused by increased storm surge and waves. As a consequence, flood risk will increase from once in a 100-year to annually for about 5 million Europeans. Perez et al. (2015) developed a statistical downscaling approach based on atmospheric conditions and past weather types. Results show a small decrease of 10 cm in wave height and period in Southern Europe.

Wahl et al. (2017) assessed flood risk and erosion rates in the Northern Gulf of Mexico. They developed a statistical model constrained by observations and based on empirical correlations for wave run up (Stockdon et al., 2006), they propagated uncertainties in the observations to changes in 100-year return period given the specific geometrical conditions. Results show a strong increase in summer up to 3 m and 0.6 m in winter. However, not everywhere storm surge and waves increases, Cannaby et al. (2015) argue that for the Singapore region trends in wave height are insignificant. Local circumstances in bathymetry and climate determine whether waves are important. [PLACEHOLDER FOR SECOND ORDER DRAFT: discuss Melet et al., in press].

4.2.3.4.3 Effects of cyclones

Tropical (TCs) and extratropical cyclones (ETCs) tend to determine extremes in sea level such as coastal storm surge, high water events, coastal flood, and their associated impacts on coastal communities around the world. The projected potential future changes in TCs and ETCs frequency, track and intensity is therefore of great importance. AR5 (WGI Chapter 14) concluded that it is *likely* that the frequency of TCs globally will either decrease or remain unchanged, but category 4/5 tropical cyclones are *likely* to increase, and also maximum wind speed and rainfall rates will increase (Christensen et al., 2013). More recently, it was realized that the modelled global frequency of TCs is underestimated and that the geographical pattern is poorly resolved in case of TC tracks, very intense TCs (i.e., category 4/5) and TC formation by using low resolution climate models (Camargo, 2013). Therefore, after AR5, multiple methods including downscaling CMIP5 climate models (Knutson et al., 2015; Yamada et al., 2017), high-resolution simulations (Camargo, 2013; Yamada et al., 2017), TC–ocean interaction (Knutson et al., 2015; Yamada et al., 2017), statistical models (Ellingwood and Lee, 2016) and statistical-deterministic models (Emanuel et al., 2008) have been developed, and the simulation capability of TCs is substantially improved. Most models still project a decrease or constant global frequency of TCs, but at the same time a robust increase in ratio of intense TCs. This is similar to IPCC AR5 and previous studies (Emanuel et al., 2008; Holland and Bruyère, 2014; Knutson et al., 2015; Kanada et al., 2017; Nakamura et al., 2017; Scoccimarro et al., 2017; Zhang et al., 2017). Through downscaling CMIP5 climate models, an increase in global TC frequency is projected during the 21st century in most locations, especially in the western North Pacific region, North Atlantic and South Indian Oceans (Emanuel, 2013).

In addition to an increase in the frequency there are also robust projected increases in the lifetimes, precipitation, and landfalls of TCs under global warming. Moreover, confidence in these projections continues to increase with the improved simulation (Walsh et al., 2016). It is *likely* that these projected increases are intensified by favourable marine environmental conditions, expansion of the tropical belt, or ocean warming in the northwest Pacific, and increasing water vapour in the atmosphere (Kossin et al., 2014; Moon et al., 2015; Cai et al., 2016; Mei and Xie, 2016; Scoccimarro et al., 2017).

Previous extensive studies also indicated the important role of warming oceans in the TC activity (Emanuel, 2005; Mann and Emanuel, 2006; Trenberth and Fasullo, 2007; Trenberth and Fasullo, 2008; Villarini and

Vecchi, 2011). Besides, TCs stir the ocean and mix the subsurface cold water to the surface, leaving a cold wake after a storm passage (Shay et al., 1992; Lin et al., 2009). Hence, ocean subsurface structure affects TC intensity. The increased thermal stratification of the upper ocean under global warming will reduce the projected intensification of TCs (Huang et al., 2015). The effect is estimated to be not more than about 15% (Emanuel, 2015; Tuleya et al., 2016). At the same time studies suggest a strengthening effect of ocean freshening in TC intensification, opposing the thermal effect (Balaguru et al., 2016). A complicating factor is that here is no physical theory to predict the number of global TCs. Currently TC frequency is broadly diagnosed using semi-empirical genesis indices, which may be problematic in predicting future global TC number (Sobel et al., 2016). We conclude that it is *likely* that the intensity of severe TCs will increase in a warmer climate, but there is still *low confidence* on the frequency change of TCs in the future.

For ETCs, AR5 concluded that the global number of ETCs is not expected to decrease by more than a few percent due to anthropogenic change. The SH storm track is projected to have a small poleward shift, but the magnitude is model dependent (Christensen et al., 2013). AR5 also found a *low confidence* in the magnitude of regional storm track changes and the impact of such changes on regional surface climate (Christensen et al., 2013). Recent projection studies indicate that trends in regional ETCs vary from region to region. Modelling studies project a significant increase in the frequency of extreme ETCs, extending from the South Atlantic across the South Indian Ocean into the Pacific (Chang, 2017). The number of storms throughout the North Atlantic basin is expected to decrease (Michaelis et al., 2017), however, an increase in the number of ETCs has been projected across the northeast North Atlantic (Colle et al., 2013; Zappa et al., 2013; Michaelis et al., 2017). The number of Mediterranean cyclones is also expected to decrease (Zappa et al., 2013). Noting that the projected frequency in ETCs still remains uncertain due to different definitions of cyclone, model biases or climate variability (Chang, 2014; Chang et al., 2016). Considering these processes imply that changes in TC and ETC characteristics will vary locally and therefore we have *low confidence* in the regional storm changes, which is in agreement with the AR5 WGI Chapter 14 (Christensen et al., 2013).

TCs and ETCs can cause storm surge, high water events, heavy precipitation, and coastal flooding. The probabilities of sea level extremes induced by TC storm surge are *very likely* to increase significantly over the 21st century due to the effect of sea level rise alone. Increasing risk from TC storm surge emerges in the highly vulnerable coastal regions, e.g., at coasts of China (Feng and Tsimplis, 2014), west Florida coast, north of Queensland, and even Persian Gulf (Lin and Emanuel, 2015; Ellingwood and Lee, 2016; Dinan, 2017; Lin and Shullman, 2017). The flood return period has greatly decreased over the past decades and is also expected to decrease greatly in the near future (by 2030–2045; Reed et al., 2015; Garner et al., 2017). For example, in New York City, the return period of a 2.25-m flood has decreased from ~500-yr before 1800 to ~25-yr during 1970–2005 and further decreases to ~5-yr by 2030–2045 (Garner et al., 2017). The annual probability of 500 mm of area-integrated rainfall induced by TC, like Harvey in 2017 in Texas, will increase from 1% in the in the late 20th century to 18% by the end of this century (Emanuel, 2017b). It is *very likely* that the flood return period in low-lying areas such as coastal megacities has decreased over the past 20th century and the high water events is expected to increase in future. In addition, the compound effects of sea level rise, storm surge and waves on extreme sea levels and the associated flood hazard are assessed in Chapter 6 (Section 6.5.3.3).

Observed damages from ETCs/TCs to coastal regions has already increased over the past 30 years and will continue in the future: an increase of 2.5°C global surface air temperature scenario is expected to increase TCs damages by 63% in the North Atlantic, and 28% in the Western North Pacific (Ranson et al., 2014). Additionally, the global population exposed to ETCs/TCs hazards has increased by almost threefold between 1970 and 2010, and this trend is expected to continue for at least a few decades (Peduzzi et al., 2012). This projected increase in coastal population will expose more people to ETCs/TCs hazards and risk in coastal regions, making preparations and evacuations more difficult and costly due to poor ETCs/TCs predictability and high-intensity landfalls (Emanuel, 2017a).

Besides, heavier precipitation and stronger low-level winds under future climate conditions are also expected for ETCs storms (Michaelis et al., 2017), and will result in increasing risk of the related damages or hazards over the North Atlantic (Michaelis et al., 2017). Onshore winds caused by ETCs can accentuate tides and enhance storm surge, resulting in severe risks at costs, e.g., battering shorelines and damaging structures (Vose et al., 2014). For example, future heavy precipitation partially induced by extreme ETCs are expected to have an impact on the timing of European floods (Blöschl et al., 2017).

4.2.3.5 Uncertainties and Decadal Predictability of Sea Level

Recent studies have explored the predictability of sea level anomalies (SLAs) out to decadal time scales, which build on similar efforts to predict sea surface temperatures and other physical variables (Meehl et al., 2009; Kirtman et al., 2013). On these time scales, SLAs typically fall well below 10 cm amplitude and are associated with changes in the ocean circulation driven by surface wind stress (e.g., Moon et al., 2013; Trenary and Han, 2013; Thompson et al., 2016), and buoyancy fluxes (Piecuch and Ponte, 2012). At high latitudes and on continental shelves, wind-driven variations in mass also contribute to SLAs (Roberts et al., 2016).

The dynamical prediction of SLA variability on seasonal to decadal time scales relies on high resolution coupled global circulation models (GCMs), initialized with the current state of the ocean-atmosphere (Kirtman et al., 2013). Miles et al. (2014) demonstrated skillful (i.e., exceeding persistence) dynamical predictions of hindcast SLA conditions up to 7 months in advance, with particularly high skill in the equatorial Pacific region. McIntosh et al. (2015) used a similar approach to examine coastal SLA predictability. Polkova et al. (2014) examined decadal hindcasts initialized every 5 years and found high predictive skill of annual-mean regional steric sea level owing to isopycnal motions (subtropics), thermosteric mixed layer changes (subtropical Atlantic), and halosteric contributions due to water mass formation (subpolar North Atlantic). Widlansky et al. (2017) combined dynamical and statistical (Chowdhury et al., 2014; Chowdhury and Chu, 2015) SLA forecasts into an operational prediction tool for Pacific sites.

The greatest uncertainty in SLA prediction is the specification of future wind conditions. Due to the complexity of the wind-forced circulation, Piecuch and Ponte (2011) contend that SLAs cannot be predicted other than variability associated with remotely forced dynamics. Likewise, SLA changes associated with the inverse barometer effect have low predictability. Examples where remotely-forced dynamics play a role, include the equatorial Pacific region associated with ENSO events (Miles et al., 2014) and at mid-latitude regions owing to westward propagating Rossby waves out to 2–5 years (Polkova et al., 2015). Other uncertainties involved in dynamical predictions include observational initialization and incomplete model physics (Kirtman et al., 2013; Hu et al., 2017).

A number of studies have linked SLA variability with climate modes (recently Hamlington et al., 2016; Han et al., 2017; Lyu et al., 2017; Moon and Song, 2017). Hence the ability to predict climate modes, e.g., decadal variability in the Pacific (Newman et al., 2016), may lead to useful predictions of related SLA patterns. As an example of the potential of this approach, Meehl et al. (2016) demonstrated that ENSO events together with long-term heat buildup or deficit in the western tropical Pacific can trigger an Interdecadal Pacific Oscillation (IPO) phase shift. Decadal predictability of the IPO also may be linked to trans-basin variability and shifts in the Walker Circulation (Chikamoto et al., 2015).

As there is no clear evidence that climate models are changing over time, we conclude with *medium confidence* that SLA magnitudes on decadal time scales will remain similar over the next century. Moreover, given the limited skills involved in predicting future wind states, we have *low confidence* in projections of SLA decadal variability.

4.2.3.6 Long-Term Scenarios, Beyond 2100

Sea level rise at the end of the century will be much higher than it is presently under most RCP scenarios. The reasons for this are mainly related to glacier melt, thermal expansion and ice sheet mass loss. These processes operate on long time scales, implying that even if the rise global temperature slows or the trend reverses, sea level will continue to rise. A study by Levermann et al. (2013) based on paleo-evidence and physical models formed the basis of the assessment by Church et al. (2013). It shows that committed sea level rise is approximately 2.3 m per degree warming for the next 2000 years. This rate is based on a relation between ocean warming and basal melt, without accounting for surface melt followed by hydrofracturing and ice cliff failure after collapse of ice shelves.

If we consider the long-term contribution of the various components of sea level rise we observe considerable differences. For glaciers, the long-term is of limited importance, because the sea level equivalent of all glaciers is restricted to ~0.4 m and there is *high confidence* that the contribution of glaciers to sea level rise expressed as a rate will decrease over the 22nd century (Marzeion et al., 2012). However, for thermal expansion and ice sheets this is not the case. For example, consideration of the effect of non-CO_2 traces cases on thermal expansion combined with the gradual rate of heat absorption in the ocean indicates that if the Montreal Protocol had been effectuated only in 2050, an additional 13 cm of sea level rise would occur in the 21st century and for more than 500 years beyond (Zickfeld et al., 2017). By far, the most important uncertainty on long timescales arises from the contribution of the major ice sheets. The time scale of response of ice sheets is thousands of years. Hence if ice sheets contribute significantly to sea level in 2100, they will necessarily also contribute to sea level in the centuries to follow. Only for low emission scenarios, like RCP2.6, can a substantial ice loss be prevented according to ice-dynamical models (Golledge et al., 2015; DeConto and Pollard, 2016).

The mechanisms for changes on these long-time scales are different for Greenland and Antarctica.

For Greenland, surface warming may lead to the condition that ablation becomes larger than accumulation, and the associated surface lowering increases ablation further in a positive feedback. As a consequence, the ice sheet will significantly retreat in the Eastern mountains. Church et al. (2013) concluded that the threshold for perpetual negative mass balance is between 1.5ºC (*low confidence*) and 4ºC (*medium confidence*) above preindustrial temperature. But one study using an intermediate complexity climate model coupled to an ice sheet model indicates a lower threshold with possible irreversibility of ice loss (Robinson et al., 2012). Passing such thresholds produces a long term contribution to sea level rise from the Greenland Ice Sheet of up to 7 m. The mechanisms for decay of the Antarctic ice sheet are related to ice shelf melt by the ocean, followed by accelerated loss of grounded ice and marine ice sheet instability, possibly exacerbated by hydrofracturing of the ice shelves and ice cliff failure (Section 4.2.3.2). The latter processes have the potential to drive faster rates of ice mass loss than the surface mass balance processes that are *likely* to dominate the future loss of ice on Greenland. Furthermore, the loss of marine-based Antarctic ice represents a long-term (millennial) commitment to elevated sea level rise, due to the long thermal memory of the ocean. Once marine based Antarctic ice is lost, local ocean temperatures will have to cool sufficiently for buttressing ice shelves to reform, allowing retreated grounding lines to readvance (DeConto and Pollard, 2016).

Several recent studies have addressed the long-term contribution of Antarctica to sea level. A minimum time scale for the Marine Ice Sheet Instability, whereby the majority of West-Antarctica decays, was derived from a schematic experiment with an ice flow model by Golledge et al. (2017), where ice shelves were removed instantaneously and prohibited from regrowing. Results of this experiment show a sea level rise from West-Antarctica of approximately 4.5 m in a century. Gradual melt of ice shelves, and partial retreat of East-Antarctic ice will lengthen this time scale to millennial or longer (Section 4.2.3.2). Prescribing a uniform warming of 2–3°C in the Southern Ocean triggers an accelerated decay of West Antarctica in a coarse resolution model with a temperature-driven basal melt formulation yielding 1 to 2 m sea level rise by the year 3000 and up to 4 m by the year 5000 (Sutter et al., 2016).

A blended statistical and physical model, calibrated by observed recent ice loss in a few basins (Ritz et al., 2015) projects an Antarctic contribution to sea level of 30 cm by 2100 and 72 cm by 2200, following a SRES A1B scenario roughly comparable to RCP4.5. The key uncertainty in these calculations was found to come from the dependency on the relation between the sliding velocity and the friction at the ice-bedrock interface. Several parameterizations are in use to describe this process. Golledge et al. (2015) present values between 0.6 m and 3 m by 2300 for the higher emission scenarios. In contrast to the previous studies Cornford et al. (2015) used an adaptive grid model, which can describe more accurately grounding line migration (Section 4.4.2). Due to the computational complexity of their model, simulations are limited to West Antarctica. Starting from present-day observations, they find that the results are critically dependent on initial conditions, sub ice-shelf melt rates, and grid resolution. The most vulnerable region was found to be the 120 km-wide Thwaites Glacier, in the Amundsen Sea sector of West Antarctica. Thwaites Glacier is currently retreating in a reverse-sloped trough up extending into the central West Antarctic Ice Sheet (Figure 4.6), where the bed is up to 2 km below sea level. The projected contributions of WAIS are found to be limited to 48 cm in 2200 following an A1B scenario. In addition to Thwaites, several smaller outlet glaciers and ice streams may contribute to sea level on long time scales, but in this study a full West-Antarctic retreat does not occur due to limited oceanic heating under the two major ice shelves (Filchner-Ronne and Ross) keeping ice streams flowing into the Ross and Weddell Seas in place. A study by DeConto and Pollard (2016) based on an ice flow model calibrated to reproduce geological sea level high-stands, shows a maximum contribution of more than 15 m sea level from the Antarctic ice sheet reached after approximately 500 years. They find the potential for considerably more sea level rise on long timescales than other studies, because they include model physics representing the influence of surface meltwater and rain on crevasse penetration (hydrofracturing of ice shelves), and the mechanical failure of ice at thick, marine terminating ice margins (marine ice cliff instability), however, the representation of these processes remains simplistic at the continental ice sheet scale (Section 4.2.3.1).

Nonetheless, recent studies using independently developed ice-dynamical models (Golledge et al., 2015; DeConto and Pollard, 2016) all agree that low emission scenarios, like RCP2.6, are required to prevent substantial future ice loss (Golledge et al., 2015; DeConto and Pollard, 2016). In ensembles of long-term simulations using a range of model physical parameters validated relative to past sea level changes, DeConto and Pollard (2016) find substantial West Antarctic ice retreat in a few RCP2.6 ensembles members, implying some risk of a substantial sea level contribution from Antarctica on century and longer timescales, regardless of the emission scenario. This is supported by observations (Rignot et al., 2014) and modelling of the Thwaites Glacier in West Antarctica (Joughin et al., 2014), suggesting grounding line retreat on the glacier's reverse sloped bedrock is already underway and possibly capable of driving major WAIS retreat on century timescales. Albeit that the driving mechanism for the retreat in those regions is ocean warming whereas the driving mechanism for retreat in DeConto and Pollard (2016) is a combination of ocean and atmospheric warming (see Section 4.2.3.1).

A study by (Clark et al., 2016) addresses the evolution of the ice sheets over the next 10,000 years and concludes that given a climate model with an equilibrium climate sensitivity of 3.5°C, the estimated combined loss of Greenland and Antarctica ranges from 25 to 52 m of equivalent sea level, depending on the emission scenario considered, with rates of GMSL as high as 2–4 m per century. A worst-case scenario was explored with an intermediate complexity climate model coupled to a dynamical ice model (Winkelmann et al., 2015), in which all readily available fossil fuels are combusted at present-day rates until they are exhausted. The associated climate warming leads to the disappearance of the entire Antarctic Ice Sheet with rates of sea level rise up to around 3 m per century.

In summary, there is *high confidence* in the continued loss of ice from both the Greenland and Antarctic ice sheets beyond 2100. A complete loss of Greenland ice contributing about 7 m to sea level over a millennium or more would occur for sustained GMST between 1°C (*low confidence*) and 4°C (*medium confidence*) above preindustrial levels. There is *low confidence* in specific estimates of the contribution of the Antarctic ice sheet beyond 2100, which range up to 15 m in 500 years, due to uncertainties regarding the dominant processes that could trigger major retreat. High-emission scenarios or exhaustion of fossil fuels over a multi-century period lead to rates of sea level rise as high as several meters/century in the long term (*low confidence*).

4.2.4 Synthesis of the Physics of Sea Level for Low-lying Islands and Coasts

This section aims to synthesize the key messages of our (geo-)physical understanding of sea level changes through time from the past to the present and future, which is important for determining exposure, vulnerability, impacts and risk related to sea level rise.

Past changes in sea level are informative as they show us the broad range of sea level in space and time over a wide range of climate conditions. These show that during past warm periods sea level was considerably higher than today. During the Eemian 130–115 Kyr BP, global temperatures are estimated to be 0.5°C–1.0°C higher than during the pre-industrial period and CO_2 around 280 ppm. For the Mid-Pliocene sea level estimates are highly uncertain but possibly up to 20 m above present-day level, temperatures 1.9 to 3.6 above present-day and CO_2 probably somewhat lower than today. Such results are a reason of concern and show that ice sheets are highly sensitive to modest warming (*high confidence*). These key finding suggest that with a few degrees of warming, substantial parts of the Antarctic and Greenland ice sheets may disappear on time scales of thousands of years or less.

However, we lack a firm understanding of the mechanisms which may lead to such an outcome and as a result, the rates of ice loss that may occur. Most modelling studies point to an insufficient contribution from Greenland during the Eemian period to explain observed sea level rise. In addition, temperature changes expected in the first half of the 21st century will not lead to a strong negative surface mass balance in Antarctica. For this reason, a lot of research focuses on which mechanisms could contribute to mass loss on the ice sheets without necessarily implying a strong further warming. Mechanisms put forward are hydrofracturing of ice shelves, Marine Ice Sheet Instability and Marine Ice Cliff Instability (Section 4.2.3.2). For these mechanisms, a small initial perturbation may induce strong positive feedbacks implying a more sensitive dynamical response of the ice sheets than observed over the last century. Geological observations provide little constraint on these processes and records of on-going changes since the start of satellite observations are too short to come to strong conclusions on possible retreat mechanisms for the ice sheet for present-day climate conditions. However, there is a growing consensus that ice-ocean interaction maybe more important than hitherto assumed (*medium confidence*).

Where Church et al. (2013) were also able to close the budget of observed sea level and the sum of the individual components, we now come to the conclusion that recent studies point to a somewhat smaller sea level rise over the 20th century implying a stronger acceleration towards its end (*medium confidence*). This does not imply that our understanding is complete: particularly at smaller local scales, which matter for society, we have difficulties explaining the observations. This is even more true where short duration events determining the frequency of extreme sea level are concerned. At the same time, the balanced budget does suggest that the major globally-acting contributors are understood. Glaciers, ice sheets and steric expansion are the key players on the global scale. With the coinciding observation of an increase in the rate of sea level rise and decrease in mass of the ice sheets, there is further awareness of the need to understand the detailed mechanism of retreat of ice sheets.

Against this background of improved understanding of the present-day rates of change, but a poor understanding of drivers of ice sheet mass changes, we evaluated projections of models of future sea level rise. As our understanding of the processes of future retreat of the Antarctic ice sheet has increased, we now include MISI like processes in the *likely* range of the projections (Section 4.2.3). The better process understanding on century time scales also implies that, especially for projection later in the 21st century, process-based models are more informative than empirical models that are based on statistical correlations over the recent past. The latter models do not implicitly capture MISI. Hence, we can rely on process-based models for the 21st century when projections within the *likely* range are sufficient for the purposes of the user community. For 2050, there is a limited scenario dependency, but for the second half of the 21st century scenarios model simulation (Golledge et al., 2015; Ritz et al., 2015; DeConto and Pollard, 2016) diverge particularly on centennial time scales as illustrated in Figure 4.10. On a millennial time-scale, the difference in GMSL between RCP2.6 and RCP8.5 is about 10 meters in some model simulations, whereas it is only decimeters at the end of 21st century and a few centimeters around 2050. Clearly, we cannot rely on process-based models to provide reliable information useful for responses to coastal risk for time scales larger than a century. Deep uncertainty remains for these time scales and probabilistic approaches are not sufficient either. This is because they are typically developed including the MICI based on the DeConto and Pollard (2016) estimates for Antarctica, which are considered to be deeply uncertain. Hence, only probabilistic scenarios (Le Cozannet et al., 2017) can be defined strongly depending on a priori assumptions for the long term processes driving the dynamics of the Antarctic ice sheet.

Whether and when this strong divergence between the scenarios will develop is impossible to judge based on existing literature and may well be convincingly shown after a tipping point is passed. Most critical in this aspect are judged to be the tipping points caused by the threshold for which ablation in Greenland gets larger than accumulation irrespective of the magnitude of the calving flux, yielding an irreversible and nearly full retreat of the ice sheet, and secondly the thresholds in ice shelf stability in West-Antarctica, depending on surface melt and sub-ice melt in combination with geometrical conditions favoring retreat. However, our ability to predict which trajectory of GMSL in Figure 4.11 is followed could occur after the tipping point is passed in time. Improved physical modelling may refine our understanding of these mechanisms and dedicated observational monitoring system will further improve our understanding. Hence, we conclude that sea level rise at the end of the century is strongly dependent on the emission scenario indicating the importance of mitigation in minimizing the risk to low-lying coastlines and islands (*high confidence*).

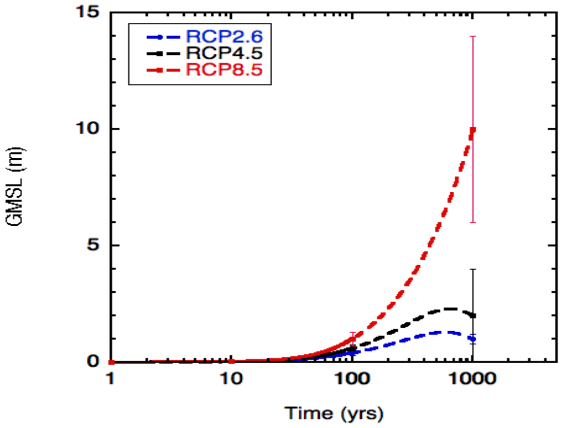


Figure 4.11: Schematic illustration of the evolution of GMSL over time, based on the *likely* range of projections for the (solid lines), indicating the large difference between different scenarios on longer than a century time scale (dotted lines) as well as the growing uncertainty both in magnitude and timing. After around 100 years, lines are dotted as they are only indicative.

Many of the impacts of sea level rise do not depend on the gradual change of sea level rise over time but rather on the combination of the trend and the many processes which are important at local scales over short periods of time. For this reason, the frequency of ESL is considered as well (Section 4.2.3). Results of the combined effect of RSL and ESL show that events which are rare (return period of 100 years or larger) in the historical context (probability < 0.01 yr^{-1}) will take place every year at some locations under each emission scenario (*high confidence*). Particularly for small islands in the Pacific which are exposed to limited variability due to storm surges, return periods will increase dramatically. Under RCP8.5 this is already the case in 2050 for most locations, whereas for lower emission scenarios this will only be the case at the end of the century (*medium confidence*).

4.3 Exposure, Vulnerability, Impacts and Risk Related to Sea Level Change

4.3.1 Introduction

Section 4.2 demonstrates that even in a stringent greenhouse gas emission scenario at the global scale (e.g., RCP2.6), GMSL has already risen and is causing detrimental effects on low-lying coastal areas. This will continue throughout the 21st Century, with significant regional and local variability, and will accelerate throughout the 21st century in case of high emission scenarios (e.g., RCP8.5). Building on the main insights from IPCC AR5, section 4.3 highlights advances in scientific knowledge about the environmental and anthropogenic drivers of exposure and vulnerability (Section 4.3.2), as well as on observed and projected impacts (Section 4.3.3). It encompasses a wide range of low-lying coastal areas, including small islands (not only Small Island Developing States), coastal cities, deltas and other continental coasts. It concludes by discussing key issues for future research on risk related to sea level rise (Section 4.3.4). Responses to projected SLR are dealt in Section 4.4.

[START BOX 4.2 HERE]

Box 4.2: Methodological Advances in Exposure and Vulnerability Assessments

Since AR5, advances have been made in exposure and vulnerability assessment to better characterize sea level rise-related coastal hazard risk, and enable the identification and localization of appropriate adaptation and risk reduction strategies although progress is very context-specific and adoption is not yet widespread (high evidence, high agreement). Exposure refers to the presence of people; livelihoods; environmental services and resources; infrastructure; or economic, social, or cultural assets in places that could be adversely affected (Glossary SREX) by sea level rise among other things. Vulnerability refers to the propensity or predisposition to be adversely affected (Glossary SREX). Main areas of post-AR5 progress include providing better projections for cascading hazards and physical impacts, such as coastal flooding and salinization, and more realistic information on exposure and vulnerability. This box showcases recent advances in assessing exposure and vulnerability to sea level rise and its physical impacts, such as coastal flooding.

Advances in exposure assessment

Many studies deal with exposure assessment, most of them considering exposure as one manifestation of risk, with a smaller number of studies interpreting exposure as a geographical location (Jurgilevich et al., 2017). Since AR5, major advances have taken place in two main areas: i) spatial-temporal assessment of exposure and ii) projected future exposure.

Improved spatial-temporal exposure assessments

The assessment of exposed elements is frequently based on census data, which is usually available at coarse resolutions. The disaggregation of census data to a higher resolution grid has often been based on proxies such as population distribution. However, technological advances (e.g., drones, mobile data, big data) and the free and ready availability of satellite data have brought, and will continue to bring, advances in exposure analysis. Exposure assessment is increasingly based on the combination of high resolution satellite imagery and spatio-temporal population modelling such as for diurnal differences in flood risk exposure (Smith et al., 2016), dynamic gridded population information for daily and seasonal differences in exposure (Renner et al., 2017), a combination of remotely-sensed and geospatial data with modelling for a gridded prediction of population density at ~100 m spatial resolution (Stevens et al., 2015), or open building data using building locations, footprint areas and heights (Figueiredo and Martina, 2016). In addition, methods based on mobile phone data (Deville et al., 2014; Ahas et al., 2015), and social media-based participation are increasingly available for population distribution mapping (Steiger et al., 2015). Some of these methodologies have been already applied in coastal assessments (e.g., Smith et al. (2016)). The level of spatial resolution is shown to impact the accuracy and precision of the risk assessment (Figueiredo and Martina, 2016) especially in case of localized hazards such as hailstorms or floods. Integrating daily and seasonal changes with the distribution of population in turn improves population exposure information for risk assessments especially in areas with highly dynamic population distributions, such as in highly touristic areas (Renner et al., 2017).

Projections of future exposure

Many climate risk and vulnerability assessments used and continue to use current data for population, land use and ecosystem data against projected future hazards (e.g., Shepherd, 2015). Recently studies have started to assess in more detail future exposure trends by accounting for the role of varying patterns of topography and development projections leading to different rates of anticipated future exposure (Kulp and Strauss, 2017), which will likely influence how effectively coastal communities might adapt (*limited evidence, medium agreement*). Other studies assess exposure considering not only projected sea levels but also expected changes in population size (Hauer et al., 2016), considering different socio-economic scenarios together with different growth rates for coastal areas and the hinterland (Neumann et al., 2015), migration-based changes in population distribution (Merkens et al., 2016; Hauer, 2017), or simulate future land use (specifically urban growth) by a cellular-automaton-based model to investigate future exposure to sea level rise (Song et al., 2017). As coastal communities may change, e.g., expand over time, the potentially exposed assets and population will change. Improvements have also been made by using spatially explicit state-and-transition simulation models for urban, residential, and rural areas (Sleeter et al., 2017) and combining them with future scenarios of risk. Most recent studies aim to account for the socio-demographic characteristics of these potentially exposed future populations in terms of their vulnerability (Shepherd, 2015); and also project future risk by future projections of socially vulnerable sub-populations (Hardy and Hauer, 2018). Using social heterogeneity modelling (Rao et al., 2017) when developing future exposure scenarios enhances the quality of risk assessments about anticipated impacts of sea level rise on coastal areas (Hardy and Hauer, 2018). Subnational population dynamics combined with an extended coastal narratives-based version of the five Shared Socioeconomic Pathways (SSP) for global coastal population distribution was used for assessing global climate impacts at the coast, highlighting regions where high coastal population growth is expected and which therefore face increased exposure to coastal flooding (Merkens et al., 2016). Relative to the year 2000, the population living in the Low Elevated Coastal Zone (LECZ) increases from 638 million to more than one billion in all SSPs by 2050. Absolute growth in exposure in the LECZ will be highest in Asia (238–303 million); Africa expects the highest relative growth (153% to 218%).

Advances in vulnerability assessment

Since the IPCC SREX report, vulnerability has been more consistently considered in climate risk assessments (*medium evidence, medium agreement*). It is recognised that climate risk is not just hazard-driven, but also a socio-economic phenomenon that evolves with changing societal conditions (*high evidence, high agreement*). Many studies related to climate risk and adaptation include vulnerability assessments, most of them considering vulnerability as a pre-existing condition while some interpret vulnerability as an outcome (Jurgilevich et al., 2017). Since AR5 major advances in the assessment of vulnerability took place in the following areas: i) understanding the importance of dynamic assessments, ii) assessing the vulnerability of social-ecological systems, iii) assessing vulnerabilities to multiple hazards simultaneously, iv) using vulnerability functions and /or thresholds instead of linear functions for more realistic outcomes, and v) using new, better data in vulnerability assessments.

Increasing importance of dynamic assessments

The dynamic nature of vulnerability, and the need to align climate forecasts with socio-economic scenarios, was a key message of IPCC SREX. Due to challenges in methodology and data availability, particularly of future socio-economic data, it is only now that an increasing number of studies include socio-economic and spatial dynamics into assessments of future vulnerability. Lack of data is overcome by downscaling global scenarios, for example, the shared socioeconomic pathways (SSPs; Van Ruijven et al., 2014; Viguié et al., 2014; Absar and Preston, 2015), or by using participatory methods, surveys and interviews to develop future scenarios (Ordóñez and Duinker, 2015; Tellman et al., 2016). The uncertainty of the downscaled projections is an issue that needs to be considered in the interpretation along with the limitation that, even if population data projections are available, the future level of education, poverty etc. is even harder to predict (Jurgilevich et al., 2017). Suggestions to overcome these shortcomings entail the use of a combination of different data sources for triangulation and inclusion of uncertainties (Hewitson et al., 2014), or the meaningful involvement of stakeholders to project plausible future socioeconomic conditions through co-production (Jurgilevich et al., 2017).

Social-ecological vulnerability assessments

The majority of existing coastal vulnerability and risk assessments focus largely on the social and/or economic dimension (Mondal, 2013; Tessler et al., 2015; Mansur et al., 2016). In many cases, especially in rural, resource-dependent settings, the relationships between people and ecosystems is a major determinant of vulnerability. Social-ecological vulnerability provides a valuable framework for identifying and understanding important social-ecological linkages, and the implications of dependencies and other feedback loops in the system. Since AR5 several methods have been developed and piloted to assess and map social-ecological vulnerability using e.g., the sustainable livelihood approach and resource dependence metrics for Australian coastal communities (Metcalf et al., 2015), integration with local climate forecasts for coral reef fisheries in Papua New Guinea (Maina et al., 2016), indicators developed in a participatory way for multiple hazards in river deltas (Hagenlocher et al., 2018), and human-nature dependencies and ecosystem services for small-scale fisheries in French Polynesia (Thiault et al., 2018). Hotspots of social vulnerability may be but are not necessarily associated with hotspots of ecosystem vulnerability, highlighting the need to specifically adapt management interventions to local social-ecological settings and to adaptation goals (Hagenlocher et al., 2018; Thiault et al., 2018). The number of social-ecological assessment studies is increasing but socio-economic factors still tend to dominate these assessments (Sebesvari et al., 2016).

Assessment of vulnerability to multiple hazards simultaneously

The same social-ecological system is often exposed to more than one hazard. Increasingly, multi-hazard risk assessments are undertaken at the coast, e.g., for erosion, flooding and inundation of coastal lands in India (Kunte et al., 2014), to understand the inter-relationships between hazards (e.g., Gill and Malamud, 2014), and by focusing on hazard interactions where one hazard triggers another or increases the probability of others occurring. Liu et al. (2016a) provide a systematic hazard interaction classification based on the geophysical environment that allows for the consideration of all possible interactions (independent, mutex, parallel, series) between different hazards, and for the calculation of the probability and magnitude of multiple interacting natural hazards occurring together. The hazard interaction classification was then piloted in China's Yangtze River Delta (Liu et al., 2016a). Also, vulnerability indicators might have to be different, depending on the hazard(s) considered. For example, the existence of shelters will lower vulnerability in the context of cyclones while it is irrelevant in case of drought. Some advances have been achieved since AR5 by using e.g. modular sets of vulnerability indicators flexibly adapting to the hazard situation (Hagenlocher et al., 2018).

Using vulnerability functions, thresholds, innovative ways of aggregation in indicator-based assessment, improved data sources

Vulnerability functions account for the fact that vulnerability and impact may not be linearly related to hazard intensity or exposure (*medium evidence, high agreement*). The use of vulnerability functions has been shown to be helpful in assessing the damage response of buildings to tsunamis (Tarbotton et al., 2015), and accounting for non-linear relationships between mortality and temperature above a 'comfort temperature' (El-Zein and Tonmoy, 2017). Several publications have shown that additive or multiplicative methodologies have weaknesses when using indicator-based vulnerability assessments (e.g., Fernandez et al., 2017). Outranking procedures and the concepts of preference, indifference and dominance thresholds have been applied as a form of data aggregation to reflect the non-compensatory nature of different vulnerability indicators (e.g. proximity to the sea cannot always be fully compensated by being wealthy) (Tonmoy and El-Zein, 2018). Similarly to advances in exposure assessments, freely available data and mobile technologies hold promise for enabling better input data for vulnerability assessments e.g. via a combination of mobile phone and satellite data to determine and monitor vulnerability indicators such as poverty (Steele et al., 2017), or to use data on subnational dependency ratios and high resolution gridded age/sex group datasets (Pezzulo et al., 2017).

[PLACEHOLDER FOR SECOND ORDER DRAFT: synopsis to be added; key messages paragraph to conclude box]

[END BOX 4.2 HERE]

4.3.2 Drivers of Exposure and Vulnerability

4.3.2.1 Key Insights from the IPCC's SREX and AR5 Publications

It is widely recognized at least since the 2012 SREX report that patterns of human development and under-development create and compound exposure and vulnerability to climate-related hazards, including SLR (*high evidence, high agreement*). Studies have progressively moved from the analysis of various parameters' influence taken individually (education, poverty, etc.,) to a more systemic approach that describes combinations of parameters, e.g., coastal urbanization and settlement patterns resulting from urban-rural discrepancies and trends in socioeconomic inequalities. The AR5 thus started differentiating between direct and indirect drivers of exposure and vulnerability (Wong et al., 2014), and between contemporary and historically-rooted drivers (e.g., trends in social systems over recent decades; Marino, 2012; Duvat et al., 2017; Fawcett et al., 2017). It also reported some progress in the development of context-specific studies, especially on coastal megacities, major deltas and small islands.

The IPCC AR5 also noted with *very high confidence* that both RSL and related impacts are influenced by a variety of local processes of social and/or environmental origin unrelated to climate (e.g., subsidence, glacial isostatic adjustment, sediment transport, coastal squeeze). Some of these processes are not clearly attributable as anthropogenic drivers, and may or may not be related to RSL, but they do influence the ability of coastal social-ecological systems to cope with and adapt to SLR and its impacts. These processes have a number of root causes and are treated here as systemic drivers that cause changes in coastal ecosystem habitat connectivity and ecosystem health conditions, for instance, and consequently the resilience of coastal livelihoods.

Very few papers, however, deal specifically the exposure and vulnerability of social-ecological systems to SLR. Rather, the literature predominantly focuses on the immediate and delayed consequences of extreme events such as tropical cyclones, storms and distant swells (see Chapter 6), for instance, and the resulting exposure and vulnerability 'in the context of SLR' (Woodruff et al., 2013). One reason for this focus is the lack, to date, of local SLR projections, given that exposure and vulnerability are very context-specific. Another reason concerns the difficulty, both for science and society, to fully comprehend long-term gradual changes like SLR (Fincher et al., 2014; Oppenheimer and Alley, 2016; Elrick-Barr et al., 2017) and ocean warming and acidification. Consequently, this section concentrates on highlighting the anthropogenic and environmental or systemic drivers that have the potential to influence exposure and vulnerability to slow-onset sea level-related hazards, thus putting aside drivers only influential in the face of extreme events.

4.3.2.2 Anthropogenic Drivers of Exposure and Vulnerability to SLR

What have we learned about anthropogenic drivers of exposure and vulnerability since AR5? And is complexity of anthropogenic drivers of exposure and vulnerability better captured?

First, emerging issues at the time of the AR5 cycle gain growing attention. This is partly due to the progressive geographical extension of social science studies dealing with climate issues, e.g., on the Arctic (Ford et al., 2012; Ford et al., 2014) and small islands (Petzold, 2016; Duvat et al., 2017), and to their downscaling at the local level, for instance, within cities (Rosenzweig and Solecki, 2014; Paterson et al., 2017; Texier-Teixeira and Edelblutte, 2017) or at the household level (Koerth et al., 2014). Due to space constrains, it is not possible here to detailed these emerging issues in an exhaustive way, and only examples can be provided. Two of them are gender inequality and the loss of indigenous and local knowledge (Cross Chapter Box 3), which more broadly reflect growing scientific and non-scientific concern about the influence of socio-economic inequalities and the decline in human-nature ties, respectively, on exposure and vulnerability to coastal hazards, including rising sea levels.

4.3.2.2.1 Gender inequality

Gender inequality, which cannot be isolated from other socio-economic dynamics, came to prominence recently in climate change studies (~15 years ago; see Pearse, 2017). In light of sea-related hazards and SLR specifically, the issue is still mainly investigated in the context of developing countries, although growing attention is paid to the situation in developed countries (e.g., Lee et al., 2015; Pearse, 2017). Recent studies in southern coastal Bangladesh, for example, show that women get less access than men to climate- and disaster-related information (both emergency information and training programmes), to decision-making processes at the household and community levels, to economic resources including financial means such as micro-credit, to land ownership, and to mobility within and outside the villages (Rahman, 2013; Alam and Rahman, 2014; Garai, 2016). Gender inequity may be inherent in unfavourable background conditions (higher illiteracy rates, deficiencies in food and calories intake, and poorer health conditions) as a result of, among other things, traditions, social norms and patriarchy. Together, these barriers disadvantage women more than men in developing effective responses to anticipate gradual environmental changes such as persistent coastal erosion, flooding and soils salinization (*medium evidence, high agreement*). Such conclusions are in line with the literature on gender inequality and climate change at large (Alston, 2013; Pearse, 2017), thus suggesting no major SLR-inherent specificities.

4.3.2.2.2 Loss of indigenous and local knowledge

Despite the identification of this issue in AR4, its treatment in AR5 was very limited, partial and ambiguous, as, for example, contradictory reference was made to indigenous people as both powerless victims in the face of climate change, and ultimate holders of valuable local knowledge to address climate change. Recent literature partly focussing on SLR reaffirms that indigenous and local knowledge (ILK) is key to determine how people recognize and respond to environmental risk (Bridges and McClatchey, 2009; Lefale, 2010; Leonard et al., 2013; Lazrus, 2015), and therefore to increase adaptive capacity and reduce long-term vulnerability (Ignatowski and Rosales, 2013; McMillen et al., 2014; Hesed and Paolisso, 2015; Janif et al., 2016; Morrison, 2017). Using examples of small islands in South-East Asia, Hiwasaki et al. (2015) describe ILK as being fully part of social-ecological processes structuring cultural traditions and activities. ILK contributes both as a foundation and an outcome to customary resource management systems aiming at regulating resources use and securing critical ecosystems protection (examples in Indonesia), at structuring the relationships between people and authorities, and at framing and maintaining a sense of the environment in the community (examples in Timor Leste). In turn, this allows local communities to predict and prepare for both sudden shock events that have historical precedent, and, when ILK is embedded in day-to-day rituals, festivals or legends, to also anticipate the consequences of gradual changes, like in sea level (examples in Indonesia). Customary resource management systems based on ILK and elders' leadership—for instance, Rahui in French Polynesia (Gharasian, 2016), or Mo in the Marshall Islands (Bridges and McClatchey, 2009)—also allow communities to diversify access to marine and terrestrial resources using seasonal calendars, to ensure collective food and water security, and to maintain ecological integrity

(McMillen et al., 2014). In rural Pacific atolls, traditional food preservation and storage (e.g., storing germinated coconuts or drying fish) still play a role in anticipating disruptions in natural resources availability (Campbell, 2015; Lazrus, 2015). Such practices have enabled the survival of isolated communities from the Arctic to tropical islands in constraining, sea-sensitive environments for centuries to millennia (McMillen et al., 2014; Nunn et al., 2017a). Morrison (2017) argues that ILK can also play a role in supporting sustainable internal migration in response to SLR, by avoiding social and cultural uprooting. It is also important to spotlight that in some specific contexts, climate change will also imply no-analogue changes, such as rapid ice-melt and changing conditions in the Arctic that have no precedent in the modern era, and could thus limit the relevance of ILK in efforts to address significantly different circumstances. Except in these specific situations, the literature suggests that the loss of ILK and related social norms and mechanisms, will increase populations' exposure and vulnerability to the impacts of SLR (Nakashima et al., 2012). The literature notably points out that modern, externally-driven socio-economic dynamics, such as the introduction of imported food (noodles, rice, canned meat and fish, etc.), diminish the cultural importance of ILK-based practices and diets locally, together with introducing dependency to monetization and external markets (Hay, 2013; Campbell, 2015). Such trends may increase long-term vulnerability to SLR (*high evidence, medium agreement*). For example, in the rural Nanumea Atoll, Tuvalu, LK supports the traditional search of 'unity or balance between the social sphere and the environmental conditions, [with the] Pre-Christian cosmology [linking] the behavior of Nanumean chiefs with the well-being of the environment' (Lazrus (2015), p. 56). In such a context, the loss of cultural ties with in situ environmental features and dynamics increases the community's exposure and vulnerability to environmental disruptions and gradual changes, notably through unsustainable livelihood practices and poor consideration of natural hazards. Finally, given that ILK is largely based on observing and 'making sense' of the environment (moon, waves, winds, animal behaviors, etc.), the loss of ILK reflects a more general concern about the loss of environmental connectedness in contemporary societies, which is not limited to remote, rural and developing communities (*medium evidence, medium agreement*). In developing contexts too, this loss of ILK has played a critical role in recent coastal disasters (e.g., Katrina in 2005 in the USA, Kates et al., 2006) and increasing vulnerability to SLR (e.g., Newton and Weichselgartner, 2014; Wong, 2014).

Second, advances have been made in understanding the complexity of anthropogenic drivers of exposure and vulnerability (Bennett et al., 2016; Duvat et al., 2017), with growing attention paid to multi-parameter, dynamic and context-specific analyses showing both the intertwining of a society's basic characteristics, and the variable direction of anthropogenic drivers' of exposure and vulnerability (Hesed and Paolisso, 2015; McCubbin et al., 2015). Accordingly, post-AR5 literature progress understanding about already-known dynamics further, some examples being described below (settlement trends, social cohesion, and risk perception).

4.3.2.2.3 Settlement trends

Major changes in coastal settlement patterns have occurred in recent decades and in the course of the 20th Century due to various socio-economic processes including population growth and demographic changes (Smith, 2011; Neumann et al., 2015), urbanization and an exodus from rural areas, tourism development, displacement and / or (re)settlement of some indigenous communities (Ford et al., 2015), changes in education levels and socio-economic disparities, etc. This has resulted in a growing number of people living in the Low Elevation Coastal Zone (~9% of the world's population; (Neumann et al., 2015; Jones and O'Neill, 2016; Merkens et al., 2016) and with significant infrastructure and assets being located in risk-prone coastal areas (*high evidence, high agreement*). High density coastal urban development is commonplace in both developed and developing countries, with extensive recent case studies, including, just to mention few, in Canada (Fawcett et al., 2017), China (Neumann et al., 2015; Jones and O'Neill, 2016; Merkens et al., 2016) and the Pacific, with ~57% of Pacific Island countries' built infrastructure located in risk-prone coastal areas (Kumar and Taylor, 2015)(*high evidence, high agreement*). Urban, high densely coastal development is a well-known illustration both in developed and developing countries, with recent case studies—just to mention few—in Canada (Fawcett et al., 2017), China (Yin et al., 2015; Lilai et al., 2016; Yan et al., 2016), Fiji (Hay, 2017), France (Yin et al., 2015; Lilai et al., 2016; Yan et al., 2016), Fiji (Hay, 2017), France (Genovese and Przyluski, 2013; Chadenas et al., 2014), Israël (Felsenstein and Lichter, 2014), Kiribati (Storey and Hunter, 2010; Duvat et al., 2013), New Zealand (Hart, 2011) and the USA (Heberger, 2012; Grifman et al., 2013; Liu et al., 2016b). [PLACEHOLDER FOR SECOND ORDER DRAFT: Synthesise and assess key findings of literature on SLR-related risk and low-lying coastal urban areas]

In Kiribati, due to the flow of outer, rural populations to limited, low-elevated capital islands, together with constrains inherent in the socio-cultural land tenure system, the built area located <20 m from the shoreline quadrupled between 1969 and 2007–2008 (Duvat et al., 2013). Population densification also affects rural areas' exposure and vulnerability. In atoll contexts, for example, the growing pressure on freshwater lenses together with a loss in LK (e.g., how to collect water from palm trees), resulted in the increased exposure of communities to brackish, polluted groundwater, inducing water security and health problems (Storey and Hunter, 2010; Lazrus, 2015). Noteworthy are other factors shaping settlement patterns, such as the fact that "indigenous peoples in multiple geographical contexts have been pushed into marginalized territories that are more sensitive to climate impacts, in turn limiting their access to food, cultural resources, traditional livelihoods and place-based knowledge [(…), and thus undermining] aspects of social–cultural resilience" (Ford et al., 2016a, p. 350). Also, "while traditional settlements on high islands in the Pacific were often located inland, the move to coastal locations was encouraged by colonial and religious authorities and more recently through the development of tourism" (Ballu et al., 2011; Nurse et al., 2014, p. 1623; Duvat et al., 2017).

4.3.2.2.4 Social capital

Recent studies confirm that besides weaknesses in the face of climate change impacts, small coastal communities also have social structures that can increase adaptive capacity to ocean-/sea-related hazards. The concept of social capital can be used in this context (Aldrich and Meyer, 2015; Petzold and Ratter, 2015). Influenced by underlying social processes, such as socioeconomic (in)equalities, gender issues, health, social networks, etc., social capital indicates the level of societal cohesion between individuals, between groups of individuals, and between people and institutions. It applies to both developing and economically advanced contexts, e.g. in densely populated deltas (Jordan, 2015), European coasts (Jones and Clark, 2014; Petzold, 2016), Asian urban or semi-urban coastal areas (Lo et al., 2015; Triyanti et al., 2017) and Pacific islands (Neef et al., 2018). It is noteworthy that social capital framed as a driver of resilience (i.e., decreasing vulnerability) is mostly studied in the context of extreme events (risk prevention mechanisms, emergency responses, post-crisis actions) and collective management of environmental features (e.g. mangroves replanting, beach cleaning, etc.), and little applied to the anticipation of gradual changes such as SLR. Some scholars, however, have started to explore the possible contribution of social capital to the public acceptability of long-term adaptation policies (Jones and Clark, 2014; Jones et al., 2015), as well as its limitations, as collective beliefs, social networks, and social and institutional trust can also negatively influence long-term vulnerability to coastal hazards (Young et al., 2014; Jordan, 2015).

4.3.2.2.5 Risk perception

Risk perception may influence communities' exposure and vulnerability as it shapes authorities' and people's attitudes towards slow-onset and/or gradual hazards—as shown by Terpstra (2011), Lazrus (2015), Elrick-Barr et al. (2017) and O'Neill et al. (2016) in case studies of the Netherlands, Tuvalu, and Australia and Ireland, respectively. For example, the deaths caused by the 2010 Xynthia storm in France resulted from demographic features (especially ageing; Vinet et al., 2012), but also from the combination of the construction of residential buildings in low-lying, flood-prone areas in recent decades; the weak maintenance of coastal dykes'; and a proportional increase in newcomers' to the region (Genovese and Przyluski, 2013; Chadenas et al., 2014). Yet, such a combination of drivers is partly rooted in progressive discounting of coastal hazard risks and subsequent loss of risk memory, as illustrated in coastal disasters such as in the aftermath of Katrina (Burby, 2006; Kates et al., 2006). Risk perception is acknowledged to be a complex anthropogenic driver of exposure and vulnerability due both to its multi-factorial nature and to its context-specific influence on policy, decision-making and action in the face of climate change (Terpstra, 2011; van der Linden, 2015). Risk perceptions stem from intertwined predictors such as "gender, political party identification, cause-knowledge, impact-knowledge, response-knowledge, holistic affect, personal experience with extreme weather events, [social norms] and biospheric value orientations" (Kellens et al., 2011; Carlton and Jacobson, 2013; Lujala et al., 2015; van der Linden, 2015, p. 112; Weber, 2016; Elrick-Barr et al., 2017). Other predictors can also come into play such as, e.g., the distance to the sea (Milfont et al., 2014; Lujala et al., 2015; O'Neill et al., 2016). Noteworthy is that there are still sometimes controversial debates on risk perception drivers. For example, knowledge about the causes and possible impacts of climate change is usually estimated a key determinant of risk perception worldwide, whether it is based on local/indigenous traditions or education levels depending on the context (Lee et al., 2015). Refining the analysis by disaggregating "knowledge" for six high-income countries, Shi et al. (2016, p. 756) show that "general scientific knowledge [is not] a robust predictor of perceived climate change risks [and that] instead, risk perceptions [are] more heavily influenced by cultural worldviews." This emphasizes that the way that "potential drivers" are measured—e.g., physical vs. perceived distance to the hazard source (O'Neill et al., 2016)—is critical to determine the nature and direction of the influence of these drivers on risk perception, and therefore of risk perception on exposure and vulnerability (Shi et al., 2016). There are however few studies on how the variability of risk perception influences exposure and vulnerability in different geographical and human contexts (e.g., Terpstra, 2011; van der Linden, 2015). There is also a critical lack of studies specifically addressing SLR. Some very recent works conducted in coastal Australia suggest that while people are confident about their ability to cope with an already experienced event, when it comes to SLR, the dominant narrative is articulated around the barriers related to the "uncertainty in the nature and scale of the impacts as well as the response options available" (Elrick-Barr et al., 2017, p. 1147). SLR is rarely addressed separately from sea-related extreme events, which masks a crucial difference between already-observed and delayed impacts. Climate change is considered as a "distant psychological risk" (Spence et al., 2012), making it and SLR per se "markedly different from the way that our ancestors have traditionally perceived threats in their local environment" (Milfont et al., 2014; Lujala et al., 2015; van der Linden, 2015, p. 112; O'Neill et al., 2016).

4.3.2.3 Environmental Drivers of Exposure and Vulnerability to SLR

Extensive conversion of coastal areas to urban, agricultural, and industrial uses exacerbates pressure on remaining ecosystems to support coastal livelihoods and to deliver ecosystem services such as storm protection, fisheries production, wildlife habitat, recreational use, tourism, and global biodiversity (Foster et al., 2017), especially in the light of sea level rise. Besides these direct anthropogenic drivers, hydrological alterations associated with climate change, including the intensity, duration, and seasonal patterns of rainfall, sea level rise, tidal range, and storm surges, also contribute to changes in the distribution and abundance of vegetation in the remaining coastal ecosystems (Gutierrez et al., 2011; Masterson et al., 2014). The degradation of coastal ecosystems contributes to both increasing human exposure and vulnerability to sea level rise (Arkema et al., 2013). On the other hand, sea level rise and its physical impacts, such as flooding or salinization, also increases ecosystem´s vulnerability and decreases the ecosystem´s ability to support livelihoods and provide coastal protection. There is a *high evidence* that healthy, diverse, connected coastal ecosystems support adaptation at the coast to sea level rise and its consequences. This section explores new knowledge since AR5 regarding processes affecting the ability of ecosystems to cope with and adapt to SLR, and associated impacts on coastal social-ecological systems and coast-dependent livelihoods, and the systemic drivers of exposure and vulnerability.

4.3.2.3.1 Recent knowledge

Large and small-scale processes influence the stability of coastal ecosystems and can interact to restrict ecosystem responses to SLR. At the large-scale, global changes in precipitation and air temperature represent a potentially significant risk to ecosystems (Garner et al., 2015; Osland et al., 2017). Maximum temperature and mean precipitation change over the last 100 years are main drivers of ecosystem stability (Mantyka-Pringle et al., 2013). In addition, seawater warming may affect marine communities and ecosystems but research remains sparse and results are contradictory (Crespo et al., 2017; Hernán et al., 2017). Also, the synergistic effects between climate change and habitat loss due to human impact and urban development are increasingly well-documented but the effects are still not well-known at larger spatial and temporal scales (Kaniewski et al., 2014; Sherwood and Greening, 2014). In addition, and although evidence is limited, recurrent disturbances may lead to losses in ecosystem adaptive capacity (Villnas et al., 2013). In summary, coastal areas and ecosystem´s responses to sea level rise around the globe is complex, with many specific responses at the ecosystem level or from keystone (foundation) species remaining poorly understood (Thompson et al., 2015) or responses are studied independently when holistic approaches may be required to understand how multiple threats affect ecosystem components, structure and functions (Giakoumi et al., 2015).

At smaller scales, risk of SLR impacts are strongly correlated with ecosystem type and species area distribution, elevation, and distance from the coast. Ecosystems and plant species that are closer to the coast, lower in elevation, and smaller in terms of their area of occurrence will be most likely to face exposure to SLR independent from species characteristics (Garner et al., 2015), although the effects are likely to be highly variable at smaller spatial scales, as shown in intertidal rocky reef habitats in Australia (Thorner et al., 2014). Even without SLR, the ecotone between coastal ecosystems and adjacent uplands responds dynamically and rapidly to inter-annual changes in inundation, with local factors, such as management of water control structures, outweighing regional ones (Wasson et al., 2013). The resulting interaction of these variables and dynamics with fragmentation, land use planning and management (Richards and Friess, 2017) has only recently started to be investigated.

Research to date has focused on identifying synergisms among stressors (Campbell and Fourqurean, 2014; Lefcheck et al., 2017; Moftakhari et al., 2017; Noto and Shurin), but interactions among them, non-linear responses, antagonisms and other feedbacks may be just as common (Brown et al., 2013; Conlisk et al., 2013; Maxwell et al., 2015; Crotty et al., 2017) and are seldom investigated, as are thresholds and tipping points in coastal ecosystem stability (O'Meara et al.; Connell et al., 2017; Wu et al., 2017). This precludes complete understanding of their complex responses (which may be greater than additive responses alone (e.g. Crotty et al. (2017)), their adequate management, or restoration regimes (Maxwell et al., 2015; Unsworth et al., 2015). Furthermore, although local management can do little to impede severe climate change impacts on ecosystems, it can slow them down and allow for evolutionary adaptation, developing alternate local management or allowing enough time to achieve the reduction of global GHG emissions necessary to slow degradation of ecosystems (Brown et al., 2013). In contrast, ecosystems with strong physical influences controlling elevation (sediment accretion and subsidence), even where mangrove replacement of salt marsh is expected, do not show changes in their vulnerability to SLR (McKee and Vervaeke, 2018), suggesting strong resilience of some natural ecosystems. In areas such as South Florida and the wider Caribbean, however, mangroves cannot outpace current SLR rates and risk disappearing. These regional and local effects are highly variable (even contradictory between studies; see Koch et al. (2015) and Smoak et al. (2013), for example) and are related to local topography and controls over salinity from freshwater and inputs (Flower et al., 2017), but further research on the mass and surface energy balance is needed (Barr et al., 2013). In addition, the responses and behavior of private landowners who may impede landward migration of ecosystems is incipient (Field et al., 2017) and, thus, highly uncertain. Overall, the long-term resilience of coastal communities and their ability to respond to rapid changes in sea level is largely unknown (Foster et al., 2017).

In addition, coastal habitat loss due to human growth and encroachment due to development, and human structures that restrict tides and, thus, interrupt mass flow processes (water, nutrients, sediments) impact tidal ecosystems depending on the type of restriction, its severity and the geomorphology of the system (Burdick & Roman 2012). The effects of coastal habitat loss are well documented (e.g. Cullen-Unsworth and Unsworth (2016); Lavery et al. (2013); Short et al. (2014); Yaakub et al. (2014); Serrano et al 2014; Breininger et al 2017), depend on the type of ecosystem and its conservation status, and interactions with SLR (Kirwan and Megonigal, 2013). Seagrass and other benthic ecosystems, for example, are declining across their range at unprecedented rates (Telesca et al., 2015; Unsworth et al., 2015; Samper-Villarreal et al., 2016) Unsworth et al 2015, Balestri et al 2017), due to degrading water quality (i.e., increased nutrient and sediment or DOC loads) from upland-based activities (which include deforestation, agriculture, aquaculture, fishing, and urbanization, port development, channel deepening, dredging and anchoring of boats (Saunders et al., 2013; Ray et al., 2014; Deudero et al., 2015; Abrams et al., 2016; Benham et al., 2016; Mayer-Pinto et al., 2016; Thorhaug et al., 2017), although the exact magnitude of area loss is still uncertain especially at smaller scales (Yaakub et al., 2014; Telesca et al., 2015). Also, human-induced impacts have facilitated the replacement of seagrasses by alternative vegetation, but the implications of habitat shifts for ecosystem attributes and processes and the services they deliver remain poorly known (Ray et al., 2014; Tuya et al., 2014). On the other hand, although threatened, coastal dunes may remain stable because their distribution is adequately covered by protected areas (Prisco et al., 2013) However, their distribution, like that of marshes and other coastal ecosystems, is limited by 'coastal squeeze', which prevents inland migration of current wetland ecosystems (Schile et al., 2014; Hopper and Meixler, 2016).

4.3.2.3.2 Coastal squeeze

Coastal squeeze was characterized in the AR5 as coastal habitat loss resulting from the combination of an eroding coastline approaching fixed and hard built or natural structures [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added]; Pontee, 2013) due to sea level rise (Doody, 2013; Pontee, 2013). The AR5 further noted that coastal squeeze is expected to accelerate due to rising sea levels [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added]. Doody (2013) characterized coastal squeeze as coastal habitats being pushed landward through the effects of sea level rise and other coastal processes on the one hand and the presence of static natural or artificial barriers effectively blocking this migration, thereby squeezing habitats in an increasingly narrower space. Pontee (2013, p. 206) clearly distinguished coastal squeeze from coastal narrowing, the latter occurring due to other forces such as changes in wind or wave patterns. Pontee (2013) further noted that it is important to understand the processes that control habitat extent and the timescales they operate at. There are therefore distinctions being made between coastal squeeze being limited to (1) the consequences of sea level rise vs. other environmental changes on the coastline and (2) the presence of only coastal defense structures vs. natural sloping land or other artificial infrastructure. Recent publications have indeed emphasised coastal squeeze related to sea level rise, although inland infrastructure blocking habitat migration was not necessarily limited to defence structures (Torio and Chmura, 2015; McDougall, 2017), and coastal ecosystem degradation by human activities leading to coastal erosion were also considered (McDougall, 2017). As long as SLR impacts remain moderate, the dominant impact will be linked to land-based development. With increased impacts of SLR the latter will be become more predominant assuming no further development on the coast.

Preserved coastal habitats can play important roles in terms of reducing risks related to some coastal hazards and initiatives are put in place to reduce coastal squeeze, such as managed realignment which includes the removal of fixed barriers inland (Doody, 2013). Coastal squeeze can lead to degradation of coastal ecosystems and species (Martínez et al., 2014), but if inland migration is unencumbered, observation data and modelling have shown that net area of coastal ecosystems could increase under various scenarios of sea level rise, depending on ecosystems considered (Torio and Chmura, 2015; Kirwan et al., 2016; Mills et al., 2016). However, recent modelling research has shown that rapid sea level rise in a context of coastal squeeze could be detrimental to the areal extent and functionality of coastal ecosystems (Mills et al., 2016) and, for marshes, could lead to a reduction of habitat complexity and loss of connectivity, thus affecting both aquatic and terrestrial organisms (Torio and Chmura, 2015). Contraction of marsh extent is also acknowledged by Kirwan et al. (2016) when artificial barriers to landward migration are in place. Adaptation to sea level rise therefore needs to account for both development and conservation objectives whereas trade-offs between protection and realignment can be found to satisfy both (Mills et al., 2016).

In summary, coastal squeeze increases a system's exposure by the loss of a buffer zone between the sea and infrastructure behind the habitat being squeezed. The clear implication is that coastal ecosystem progressively lose their ability to provide regulating services with respect to coastal hazards, including with respect to the risk posed by sea level rise in terms of inundation and salinization. The vulnerability of communities is increased through the loss of other ecosystem services that these ecosystems could provide, for example in terms of direct income linked to tourism or when livelihoods are directly dependent on these ecosystems. Vulnerability is also increased when freshwater resources become salinized, particularly in the case when these resources are already scarce.

4.3.2.3.3 Subsidence

Subsidence in coastal zones, in general, and in coastal deltas, in particular, poses serious development and adaptation challenges with respect to sea level rise. In many cases, the rates of natural land subsidence are accelerated because of human activities, such as extraction of natural resources e.g., water, oil, and gas (AR4; PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added]). These rates can often be much higher than eustatic sea level rise. For example, Higgins et al. (2014) showed that for the Ganges-Brahmaputra delta in Bangladesh subsidence rates ranging from 0 to >18 mm yr^{-1} are recorded depending on location and local stratigraphy. A review covering the Ganges–Brahmaputra–Meghna indicates great variability in subsidence (or uplift) rates but reports that average annual subsidence rates during the last 1,000 years (at 8.8 mm yr^{-1} but with high variability) was four times faster than for prior periods, although the difference could be attributed to e.g., differences in measurement methods, among other factors (Brown and Nicholls, 2015). Similarly, in the Mekong delta portion of Vietnam, modelled groundwater extraction related subsidence covering the past 25 years has shown an average of approximately 18 cm of subsidence over the region, with areas in excess of 30 cm subsidence, with present average annual rates of 1.1 cm yr^{-1} and highs of about 7 cm yr^{-1} in Ho Chi Minh City (Minderhoud et al., 2016).

In summary, land subsidence increases exposure to sea level rise through the loss of elevation, allowing seawater to creep inland at a faster pace. In many cases, land subsidence proceeds at a rate that is faster than eustatic sea level rise, thus compounding the effects of the latter. The consequences are higher exposure to both inundation and salinization of resources. As for coastal squeeze, these factors also contribute to increased vulnerability directly (particularly when sources of freshwater are scarce), and indirectly (as often, land subsidence is due in part to groundwater over-abstraction which, when the resource becomes depleted, polluted or salinized, limits is usage for economic activities and domestic consumption).

4.3.2.3.4 Catchment connectivity, upstream effects

Coastal areas, including deltas, are highly dynamic as they are affected by natural and/or human-induced processes locally or originating from both the land and the sea. Rapid changes taking place in the catchment can therefore have severe consequences for coastal areas in terms of e.g., pollution, erosion, and/or land subsidence. A critical factor is the sediment supply reaching the coast (Tessler et al., 2018). For instance, Anthony E.J. (2015) reported substantial erosion in the Mekong delta between 2003 and 2012 which was attributed in part to a reduction in surface-suspended sediments in the Mekong river potentially linked to dam construction within the river basin, sand mining in the river channels, and land subsidence linked to groundwater over-abstraction locally. Schmitt et al. (2017) demonstrated that these and other drivers in sediment budget changes can have severe effects on the very physical existence of the Mekong delta by the end of this century, with the most important single driver leading to inundation of large portions of the delta being ground-water pumping induced land subsidence. Another, rarely considered factor is the shift in tropical cyclone climatology which also plays a critical role in explaining changes in fluvial suspended sediment loads to deltas as demonstrated by Darby et al. (2016), again for the Mekong delta. More generally, most conventional engineering strategies that are commonly employed to reduce flood risk (including levies, sea-walls, and dams) disrupt a delta's natural mechanism for building land. These approaches are rather short-term solutions which overall reduce the long-term resilience of deltas (Tessler et al., 2015; Welch et al., 2017). Systems particularly prone to flood risk due to anthropogenic activities include North America's Mississippi River delta, Europe's Rhine River delta, and deltas in East Asia (Renaud et al., 2013; Day et al., 2016). In regions where suspended sediments are still available in relatively large quantities, rates of sedimentation during flooding seasons can vary depending on multiple factors, including the type of infrastructure present locally, as was shown by Rogers and Overeem (2017) for the Ganges-Brahmaputra-Meghna (Bengal) Delta in Bangladesh. Overall, reduced freshwater and sediment inputs from the river basins are critical factors determining delta sustainability (Day et al., 2016; Renaud et al., 2013). In some contexts, this can be addressed through basin-scale management which allow more natural flows of water and sediments through the system, including methods for long-term flood mitigation such as improved river-floodplain connectivity, the controlled redirection of a river (i.e., avulsions) during times of elevated sediment loads, the removal of levees, and the redirection of future development to lands less prone to extreme flooding (Renaud et al., 2013; Day et al., 2016; Brakenridge et al., 2017). These actions could potentially increase the persistence of coastal landforms in the context of sea level rise.

In summary, catchment-scale changes have very direct impacts on the coastline, particularly in terms of water and sediment budgets (*high confidence*). The changes can be rapid and modify coastlines over short periods of time, outpacing the effects of SLR leading to increased exposure and vulnerability of social-ecological systems (*high confidence*). Towards the end of the century, SLR may even generate greater impacts. Without losing sight this fact, it is however imperative that catchment-level processes be understood and managed to limit rapid increases in exposure and vulnerability.

4.3.2.4 Towards a Synthetic Understanding of the Drivers of Exposure and Vulnerability

Recent literature irrevocably confirms that anthropogenic drivers played a major role, over the last century, in the increase of exposure and vulnerability worldwide and they will continue to do so in the absence of adaptation (*high evidence, high agreement*). Some scholars argue that '(…) even with pervasive and extensive environmental change associated with ~2°C warming, it is non-climatic factors that primarily determine impacts, response options and barriers to adapting' (Ford et al. (2015), p. 1046). Although it is the interaction of climate and non-climate factors that eventually determine the level of impacts, such awareness has important implications for action, especially by showing that major action can be undertaken already in favor of long-term adaptation and despite uncertainty of local climate change impacts (Magnan et al., 2016a). Similarly, the state and condition of coastal ecosystems largely influence their capacity to cope with or adapt to sea level rise and it impacts.

We now better understand the diversity and interactions of the climate and non-climate drivers of exposure and vulnerability. As a result, we realize how much context-specificities (geography, social inequity, risk perceptions etc.) play a critical role in shaping the direction of influence of individual drivers and of their possible combinations on the ground (*high evidence, high agreement*).

Last, recent studies (e.g., cited in this section) also confirm AR5 conclusions that both developing and developed countries are exposed and vulnerable to SLR (*medium evidence, medium agreement*).

The ability of coastal ecosystems to serve as a buffer zone between the sea and human settlements of infrastructure, and to provide regulating services with respect to coastal hazards, including to sea level rise in terms of inundation and salinization, is progressively being lost due to coastal squeeze, pollution, habitat degradation etc.; ecosystem degradation being one driver of exposure and vulnerability in the coastal zone (*high evidence, high agreement*).

4.3.3 Observed Impacts, and Current and Future Risk of SLR

Climate change induces modifications to the ocean's and the cryosphere's physical and chemical parameters (ice density, permafrost thaw rates, river water flows, ocean pH, sea-surface temperature, etc.), which explain sea level rise (SLR; section 4.2). SLR will then combine with extreme events (e.g., storms) to generate coastal hazards (marine flooding, coastal erosion, etc.), and in turn consequences on ecosystems (marshes and mangroves, coral reefs, seagrass), natural resources (e.g., groundwater lenses) and ecosystem services (e.g., coastal protection). Together with the influence of anthropogenic drivers (section 4.3.2.2), this results in direct and indirect impacts on human systems, e.g., people/assets/infrastructures exposed, agriculture, tourism, fisheries and aquaculture, socioeconomic inequity and well-being, etc. Figure 4.12 gives an overview of potential SLR-induced effects.

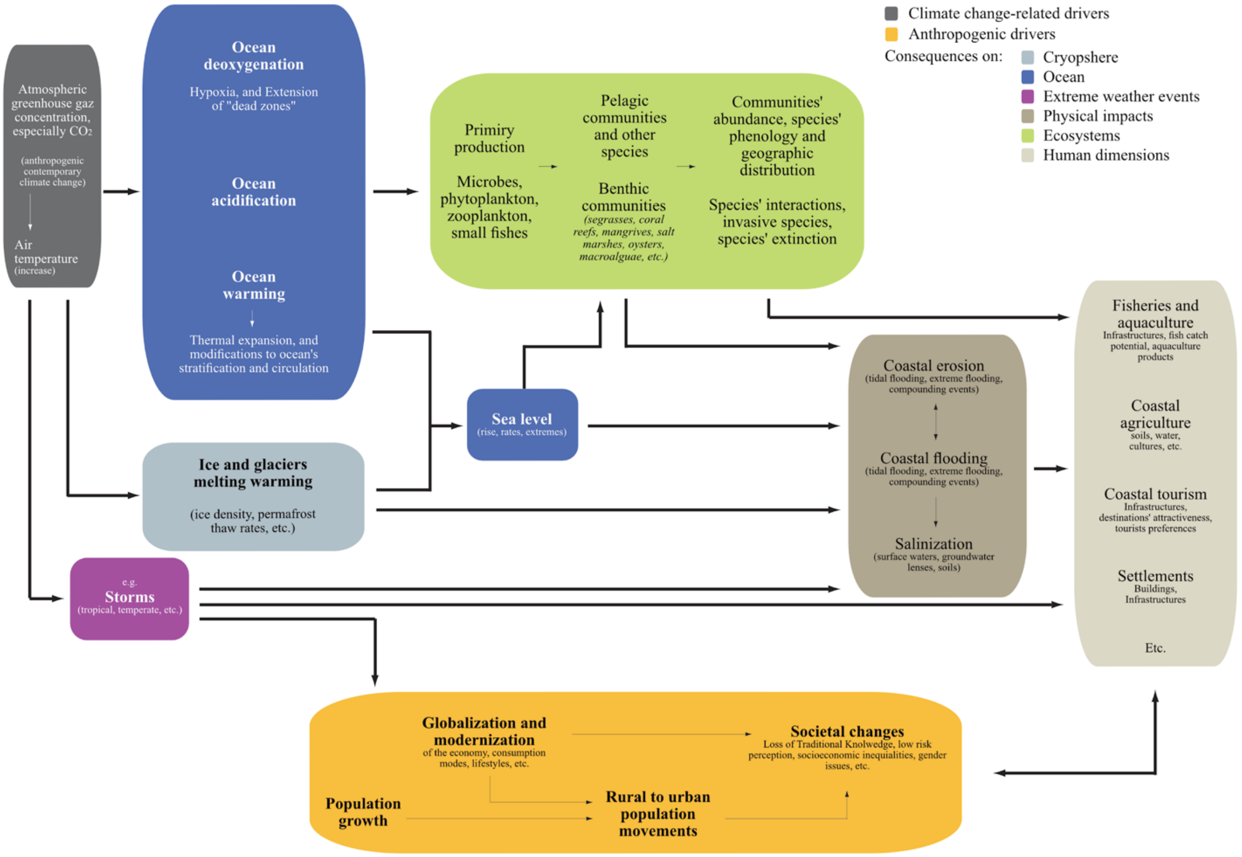


Figure 4.12: Overview of the main cascading effects of sea level rise. This figure presents a generic, theoretical understanding of the cascading effects to be expected from SLR, including other climate change-related changes in the ocean's and cryosphere's physical and chemical parameters (blue, blue-grey boxes) and extreme events (purple box; storms are used as an example). It shows the implications for marine and coastal ecosystems (green boxes) as well as in

terms of coastal hazards (erosion, flooding, salinization; dark brown box); and eventually for human dimensions (key economic sectors and settlements; light brown box). It also shows the role played by anthropogenic drivers in generating risk (orange box). No specific climate change scenario is considered. This Figure builds on the main findings of the IPCC AR5 (Larsen et al., 2014; Nurse et al., 2014; Wong, 2014), as well as of this IPCC Special Report [Chapters 1 to 6].

4.3.3.1 Observed and Projected Physical Impacts

Three major concerns for low-lying coasts are marine flooding, coastal erosion and salinization (Seneviratne et al., 2012; Nurse et al., 2014; Wong, 2014), which can be temporary (e.g., due to a storm event, or to seasonal variability in sediment transport), or permanent. Other processes also influence permanent changes especially, such as, e.g., starvation of sediments provided by rivers (Kondolf et al., 2014); permafrost thaw and ice retreat (Cramer et al., 2014); or the disruption of natural dynamics by coastal development and activities such as land reclamation or sediment mining. The sections below describe projected impacts assuming no adaptation.

4.3.3.1.1 Marine flooding

Marine flooding results from the combination of multiple parameters, such as storms (wind speed, intensity, size, angle approach, etc.), distant-source swells (Cooper et al., 2013; Hoeke et al., 2013; Smithers and Hoeke, 2014), landfall location (size, topography, etc.) and coastal development (urban building practices, vegetation removal, anthropogenic-driven subsidence, etc.). In addition to be temporary or permanent, recent literature highlights that marine flooding can also be chronic, i.e., when high tides occur under calm weather conditions (Sweet and Park, 2014; Moftakhari et al., 2015; Dahl et al., 2017).

Marine flooding is already affecting deltas around the world; ~260,000 km^2 have been temporarily submerged over the 1990s/2000s (Syvitski et al., 2009; Wong, 2014). Because coastal flooding is driven by many socio-economic, ecological, and physical factors, it is difficult to make projections on purely physical impacts. However, models are improving to account for the morphological diversity of coasts and variability of hydrodynamic forcing (e.g., Pearson et al. (2017)), and some combine SLR and storm surge events. For example, Vitousek et al. (2017)(p. 6) estimate that 'only 5-10 cm of SLR, expected under most projections to occur between 2030 and 2050, doubles the flooding frequency in many regions, particularly in the Tropics'. A larger rise in sea level of 1.2 m at the end of the century is estimated to multiply by ~2 to 5 the flooded areas for coastal communities along the east coast of the US (Dahl et al., 2017). Also using a 1.2 m SLR scenario by the end of the century, Lilai et al. (2016) estimate that 0.47 % of the Xiamen city area, China, will be permanently inundated, a proportion increasing to 10.5 % when also including a 200-year return period storm tide. However, regional responses will differ, due notably to contrasting changes in extreme wave energy fluxes along the northern (decrease) and southern (increase) hemispheres' shorelines (Mentaschi et al., 2017), and in reef ecosystem response to climate-related stressors (Beetham et al., 2017).

4.3.3.1.2 Coastal erosion

While coastal erosion is a well-known problem, quantified assessments of its significance worldwide are still lacking (Cazenave and Cozannet, 2014). Recent literature suggest that the phenomenon is expanding in many regions, e.g., Brazil (Amaro et al., 2015), China (Yang et al., 2017), Colombia (Rangel-Buitrago et al., 2015), the Arctic (Mars and Houseknecht, 2007; Jones et al., 2009), and the western Pacific (Albert et al., 2016; Garcin et al., 2016). Since the AR5, however, there is growing appreciation for and understanding of the ability of coastal systems to respond dynamically to SLR (Passeri et al., 2015; Lentz et al., 2016; Deng et al., 2017). Most low-lying coastal systems exhibit important feedbacks between biological and physical processes that have allowed them to maintain a relatively stable morphology under moderate rates of SLR ($<$ 0.3 cm yr^{-1}) over the past few millennia (Woodruff et al., 2013; Cross Chapter Box 2).In a global review on multi-decadal changes in the surface area of 709 atoll reef islands, Duvat (Submitted) shows that in a context of more rapid SLR than the global mean (Becker et al., 2012), 73.1% of islands were stable in area, while respectively 15.5 % and 11.4 % increased and decreased in size. This suggests that some low-lying coastal systems have had the capacity to naturally adjust to SLR until now. However, this capacity is expected to be reduced in the coming decades, due to the combination of high SLR rates, increased wave energy (Albert et al., 2016), changes in run-up (Shope et al., 2017) and storm wave direction (Harley et al., 2017), ocean warming and acidification, and an expected increase in anthropogenic pressure.

4.3.3.1.3 Salinization

Salinization describes the consequences of saline or brackish water intrusion both by submergence of the surface and by ground penetration in the case of porous soils made of sand or alluvium, for example. In river deltas and estuaries, where saline or brackish water can enter landwards with the tide easily, salinization may impact ecosystems, water supply and livelihoods far inland. Saline and brackish water intrusion has a pronounced impact on both ecosystems and social systems by increasing salinity levels of groundwater, surface water, and soils with associated shortages of freshwater resources and challenges for coastal ecosystems and livelihoods (*high evidence, high agreement*).

Coastal groundwater lenses

Groundwater volumes will primarily be affected by variations in precipitation patterns (Taylor et al., 2013; Jiménez Cisneros et al., 2014), which are expected to increase water stress in small islands (Holding et al., 2016). While SLR will mostly impact groundwater quality (Bailey et al., 2016), exacerbating marine flooding events-induced salinization (Gingerich et al., 2017), it will also affect the water-table height (Rotzoll and Fletcher, 2013; Jiménez Cisneros et al., 2014; Masterson et al., 2014; Werner et al., 2017) and barrier islands (Masterson et al., 2014). This will have consequences on both freshwater availability (for people and agriculture) and vegetation dynamics. At many locations, direct anthropogenic influences such as groundwater pumping for agricultural or urban uses, impact salinization of coastal aquifers more strongly than SLR in the 21st century (Ferguson and Gleeson, 2012; Jiménez Cisneros et al., 2014; Uddameri et al., 2014), with trade-offs in terms of groundwater depletion that may contribute to land subsidence and thus increase marine flooding risk. The natural migration of groundwater lenses inland in response to SLR can also be severely constrained by urbanization, e.g. in semi-arid South Texas, USA (Uddameri et al., 2014). Yet, the influence of land-surface inundation on seawater intrusion and resulting groundwater lenses salinization has been underestimated until now (Ataie-Ashtiani et al., 2013; Ketabchi et al., 2014). Such sea-borne impacts will potentially also combine with a projected drying of most of the tropical-to-temperate islands by mid-century (Karnauskas et al., 2016).

Surface waters

The quality of surface water resources (in estuaries, rivers, reservoirs, etc.) can be affected by the intrusion of saline and/or brackish water, both in a direct (increased salinity) and indirect way (altered environmental conditions which change the behavior of pollutants and microbes). In terms of direct impacts, statistical models and long-term (from 1950 to present) records of salinity show significant upward trends in salinity and a positive correlation between rising sea levels and increasing residual salinity, e.g., in the Delaware Estuary, USA (Ross et al., 2015). Higher salinity levels, further inland, have also been reported in the Gorai river basin, Southwestern Bangladesh (Bhuiyan and Dutta, 2012), and in the Mekong Delta, Vietnam. In the latter, salinity intrusion extends ca.15 km inland during the rainy season and typically ca. 50 km during dry season (Gugliotta et al., 2017). Brackish water species such as mangroves, mollusks, and diatoms, have been reported even more inland, demonstrating that in the Mekong Delta, low salinity brackish water may reach up to 160 km inland during the dry season. More broadly, the impact of salinity intrusion can be significant in river deltas or low-lying wetlands, especially during low-flow events such as in the dry season (Dessu et al., 2018). In Bangladesh, e.g., some freshwater fish species are expected to lose their habitat with increasing salinity, with profound consequences on fish-dependent communities (Dasgupta et al., 2017). In the Florida Coastal Everglades, sea level increasingly exceeds ground surface elevation at the most downstream freshwater sites, affecting marine-to-freshwater hydrologic connectivity and transport of salinity and phosphorous upstream from the Gulf of Mexico. In the Everglades, the impact of SLR is higher in the dry season when there is practically no freshwater inflow (Dessu et al., 2018), and salinity intrusion was shown to also cause shifts in the diatom assemblages, with expected cascading effects through the ecosystem and the food web (Viviana Mazzei, 2018). Further impacts include limitations in drinking water supply due to salinization (Wilbers et al., 2014), projected fresh water shortage in reservoirs, e.g., for Shanghai (Li et al., 2015a). Salinity intrusion also indirectly affects surface water quality. Salinity changes the partitioning and mobility of some metals, and hence of their concentration or speciation in the water bodies (Noh et al., 2013; Wong et al., 2015; de Souza Machado et al., 2018). Varying levels of salinity also influence the abundance and toxicity of *Vibrio cholerae* in the Ganges Delta (Batabyal et al., 2016).

Soils

Soil salinisation is one of the major soil degradation threats, with sea water intrusion being one of the common causes (Daliakopoulos et al., 2016). Sea water intrusion leads to the salinization of exposed soils with changing carbon dynamics (Ruiz-Fernández et al., 2018) and microbial communities (Sánchez-Rodríguez et al., 2017), impacting soil enzyme activity (Zheng et al., 2017), metal toxicity (Zheng et al., 2017), plant germination (Sánchez-García et al., 2017), biomass production (Yao et al., 2015), yield (Genua-Olmedo et al., 2016), and also soil-born greenhouse gas production (Liu et al., 2017; Sánchez-Rodríguez et al., 2017; Ruiz-Fernández et al., 2018). In a study in the Ebro Delta, Italy, soil salinity was shown to be directly related to distances to the river, to the delta inner border, and to the river old mouth (Genua-Olmedo et al., 2016). Land elevation was the most important variable in explaining soil salinity. Sea level rise was shown to decrease C_{org} concentrations and stocks in sediments of salt marshes as reworked marine particles contribute with a lower amount of C_{org} than terrigenous sediments. C_{org} accumulation in tropical salt marshes can be as high as in mangroves and the reduction of C_{org} stocks by ongoing sea level rise might cause high CO_2 releases (Ruiz-Fernández et al., 2018). Pore water salinity levels in coastal marsh soils can become significantly elevated in just one week of flooding by sea water, which can potentially negatively impact macrophytes and associated microbial communities for significantly longer time periods (McKee et al., 2016). Sea level rise will also alter the frequency and magnitude of wet/dry periods and salinity levels in coastal ecosystems, with consequences on the formation of climate-relevant greenhouse gases, such as CH_4, CO_2, and N_2O (Liu et al., 2017).

4.3.3.1.4 Attribution of observed physical changes to SLR

The AR5 concludes on the major difficulty to attribute observed changes to SLR per se because 'the coastal sea level change signal is often small when compared to other processes' (2014 Wong (2014), p. 375). On coastal morphological changes, e.g., contemporary SLR currently acts as a "background natural driver", with extreme events, changes in wave patterns and wave energy fluxes along shorelines, and human intervention often described as the prevailing drivers of observed changes. While recent work confirms the complexity of the attribution issue (e.g., Romine et al. (2013); Le Cozannet et al. (2014)), others bring new insights on possibly emerging signs of recent SLR influence on shoreline dynamics, especially on low-lying, sensitive coasts in New Caledonia (Garcin et al., 2016), the Federal States of Micronesia (Nunn et al., 2017b) and the Solomon Islands (Albert et al., 2016). Early signs of recent SLR direct influence on estuaries' water salinity are also emerging, e.g., in the Delaware, USA, where Ross et al. (2015) estimate a salinity increase by as much as 4.4 psu (= g Na+Cl- per Liter) per meter of SLR. Overall, while the literature suggests that it is still too early to attribute coastal impacts to SLR in most of the world's coastal areas, there is some agreement that as SLR will continue rising, the frequency, severity and duration of hazards and related impacts, will possibly increase (Woodruff et al., 2013; Lilai et al., 2016; Vitousek et al., 2017), and as soon as the second half of the 21st century for impacts on shoreline dynamics (Storlazzi, 2018).

4.3.3.2 Observed and Projected Impacts on Ecosystems and Ecosystem Services

Ecosystems provide natural resources and various services, e.g., cultural (e.g., recreation and ecotourism, or aesthetic values) and protection against sea-related hazards (e.g., waves). Due to space constraints, this section only discusses some examples of critical marine ecosystems (marshes, mangroves, coral reefs and seagrasses) and ecosystem services (coastal protection), although it is recognized that terrestrial ecosystems (e.g., sand dune vegetation) also play an important role.

4.3.3.2.1 Marshes and mangroves

Potentially one of the most important of the eco-morphodynamic feedbacks allowing for relatively stable morphology under SLR, is the ability of marsh and mangrove systems to enhance the trapping of sediment, which in turn allows wetlands to grow, and increase the production and accumulation of organic material (Kirwan and Megonigal, 2013). When ecosystem health is maintained and sufficient sediment exists to support accretion, this particular feedback has generally allowed marshes and mangrove systems to build vertically at rates equal to or greater than SLR up to present day (Kirwan et al., 2016; Woodroffe et al., 2016). While recent reviews suggest that mangroves' surface accretion rate will only keep pace with high SLR scenario (RCP8.5) up to years 2055 and 2070 in fringe and basin mangrove settings, respectively (Sasmito et al., 2016), process-based models of vertical marsh growth that incorporate biological and physical feedbacks rather support survival under rates of SLR as high as 1-to-5 cm/yr before drowning (Kirwan et al., 2016). These rates are substantially higher than what could be supported without vegetation and highlight the importance of eco-morphodynamic feedbacks for maintaining and building new land at the coast. Threshold rates of SLR before marsh drowning however vary significantly from site-to-site and can be substantially lower than 1 cm.yr^{-1} in micro-tidal regions where the tidal trapping of sediment is reduced and/or in areas with low sediment availability (Lovelock et al., 2015; Ganju et al., 2017; Watson et al., 2017). In the extreme case of no sediment supply, it has been estimated that salt marshes cannot accrete faster than 3 mm.yr^{-1}, and clastic sediment supply may limit many wetlands along the Gulf and East Coast of the US to a threshold SLR of less than 0.5 cm/yr (Morris et al., 2016). Processes impacting lateral erosion are just as, if not more important, than vertical accretion rates in determining coastal wetland survival (e.g., Mariotti and Carr (2014)). In general most marsh and mangrove systems established themselves at their current locations over the last few thousand years and under relatively slow rates of sea level change (Newman, 1965; Redfield, 1972; Ellison, 1991; Parkinson, 1994). Preserved marsh peat that dates prior to this interval, when rates of SLR were similar or greater than Present, are predominantly found along open-beach faces and off-shore, indicating that these systems were not stationary, but instead migrating landward or transgressing under these higher rates of SLR (Kirwan and Megonigal, 2013). However, these off-shore records are incomplete due to the erosive nature of this landward migration making it difficult to assess how prominent marsh and mangrove environments were along the coast during earlier epoch of rapid shore-line retreat (Parkinson, 1994). For most low-lying coastlines, a seaward loss of wetland area due to marsh retreat could be offset by a similar landward migration of coastal wetlands (Kirwan and Megonigal, 2013; Schile et al., 2014) [PLACEHOLDER FOR SECOND ORDER DRAFT: IPCC SR1.5]), this landward migration having the potential to maintain and even increase the extent of coastal wetlands globally (Morris et al., 2012; Kirwan et al., 2016). This natural process will however be constrained in areas with steep topography or equipped with hard engineering structures (i.e., coastal squeeze). Seawalls, levees and dams can also prevent the fluvial and marine transport of sediment to wetland areas and reduce their resilience further (Giosan, 2014; Tessler et al., 2015; Day et al., 2016; Spencer et al., 2016).

4.3.3.2.2 Coral reefs

Coral reefs have recently become iconic symbols of the threat of climate-related ocean change, especially ocean warming and acidification, even under a RCP2.6 scenario, to ecosystems and communities (Hoegh-Guldberg, 2014; Gattuso et al., 2015; Albright et al., 2018). For example, the 2016 coral bleaching event caused extensive coral mortality especially in the Pacific and Indian oceans (Hughes et al., 2017; Perry and Morgan, 2017), and 'more than half of the world's reefs are under medium or high risk of degradation' (Gattuso et al. (2014), p. 97). In sharp contrast to the susceptibility of coral reefs to ocean warming and acidification, some studies suggest that SLR is *likely* to have negligible impacts on coral reefs' vertical growth because the projected rate and magnitude of SLR by 2100 are within the potential accretion rates of most coral reefs (van Woesik et al., 2015). Other scholars, however, stress that the overall net vertical accretion of reefs may decrease after the first 30 years of rise in a 1.2 m SLR-scenario (Hamylton et al., 2014). The AR5 concludes that 'a number of coral reefs could (...) keep up with the maximum rate of sea level rise of 15.1 mm yr^{-1} projected for the end of the century (*medium confidence*) but a lower net accretion than during the Holocene (Perry et al., 2013) and increased turbidity (Storlazzi et al., 2011) will weaken this capability (*very high confidence*)' (Wong (2014), p. 379). A key point is that SLR will indeed not act alone. The cumulative impacts of other natural (notably sea surface warming) and anthropogenic drivers are estimated to reduce the ability of coral reefs to keep pace with future SLR (Hughes et al., 2017; Yates et al., 2017) and thereby continue providing sediments and protection to coastal areas. Ocean acidification is estimated to slow growth rates and reef accretion (e.g., Eyre et al. (2018), Albright et al. (2018)) and some have suggested coral's vertical growth may be unable to keep pace with projected SLR over the 21st century due to ocean warming (Perry and Morgan, 2017). Some studies, e.g., in Palau (van Woesik et al., 2015), are more optimistic as they conclude that coral reefs will keep growing vertically in the case of stringent greenhouse gas emission mitigation scenarios (especially RCP2.6). However, the global society' ability to be on track to RCP2.6 is still far from certain, thus balancing the Woesik et al.'s conclusion. Another concern is that locally, even small SLR can increase turbidity on fringing reefs through increased resuspension of fine sediment on reef flats and increased coastal erosion and transport of fine sediment to adjacent reefs. This will potentially interfere with photosynthesis, feeding, recruitment, and other key physiological reef processes (Field et al., 2011; Siegle and Costa, 2017).

4.3.3.2.3 Seagrasses

Due to their natural capacity to enhance accretion and in the absence of mechanical or chemical destruction by human activities, seagrass are not expected to be severely affected by SLR per se, except indirectly through the increase in extreme weather events' and waves' on coastal morphology (i.e., erosion) and though changes in light levels (and although sometimes mediated through effects on adjacent ecosystems; Saunders et al. (2013)). Extreme flooding events have also been shown to cause large-scale losses of seagrass habitats

(Bandeira and Gell, 2003), and seagrasses in Queensland, Australia, were lost in a disastrous flooding event (Campbell and McKenzie, 2004). Changing current patterns can also either erode seagrass beds or create new areas for seagrass colonization. But overall, seagrass will primarily be negatively affected by the direct effects of increased sea temperature on growth rates and the occurrence of disease (Marba and Duarte, 2010; Chapter 5; Burge et al., 2013; Koch et al., 2013; Thompson et al., 2015; Gattuso et al., Submitted; Chapter 5), and by heavy rains that may dilute the seawater to a lower salinity. Noteworthy is that some positive impacts are expected, e.g., as ocean acidification is *likely* to benefit photosynthesis and growth rates of seagrass (Repolho et al., 2017).

4.3.3.2.4 Coastal protection by marine ecosystems

Major 'protection' benefits derived from the above-mentioned coastal ecosystems include wave attenuation and shoreline stabilization, e.g., by coral reefs (Elliff and Silva, 2017; Siegle and Costa, 2017). However, climate change and specifically SLR impacts, may reduce these ecosystem services. Recently, a global meta-analysis demonstrated that, on average, these ecosystems together reduce wave heights between 35%–71% (Narayan et al., 2016), with coral reefs, salt-marshes, mangroves and seagrass/kelp beds reducing wave heights by 70%, 72%, 31% and 36%, respectively. Global analyses show that natural and artificial seagrasses can attenuate wave height and energy by as much as 40% and 50%, respectively (Fonseca and Cahalan, 1992; John et al., 2015), while coral reefs reduce wave energy by an average of 97% (Ferrario, 2014) and wave-driven flooding volume by 72% (Beetham et al., 2017). In addition, it is noteworthy that the effectiveness of marshes and reefs in attenuating waves is unrelated to their effectiveness in attenuating surge (i.e., wave heights could be reduced but not necessarily the elevation of inundation (Shepard CC, 2011; Brandon et al., 2016; Castagno, In review). Other ecosystems provide coastal protection, including macroalgae, oyster and mussel beds, and also beaches, dunes and barrier islands, but there is less understanding of the level of protection conferred by these other organisms and habitats (Spalding et al., 2014). Additionally, human-driven pressure on these ecosystems is inherently difficult to forecast due to the possible implementation of new policies and the effectiveness of management and climate mitigation efforts.

4.3.3.3 Observed and Projected Impacts on Human Systems

The section on impacts of sea level rise on human settlement in AR5 concluded that only a small fraction of the underlying climate uncertainty has been explored. Specifically, the following aspects have been under researched: (i) impacts of high-end SLR, (ii) impacts of regional patterns of climate-induced sea level and anthropogenic subsidence, (iii) impacts under various urbanization scenarios for example, (iv) impacts under a range of plausible adaptation scenarios considering a wide range of adaptation measures (Wong, 2014). Since AR5 progress has been made on some of these aspects, specifically for the SLR impact of enhanced coastal flooding.

4.3.3.3.1 Coastal flood risks

Global assessments of current and future coastal flood exposure and risk have explored a much wider range of uncertainties than considered in AR5. Exposure studies accounted for subnational human dynamics such as coastward migration or coastal urbanization, which increases estimates of the population living in the Low Elevation Coastal Zone (LECZ) in 2100 by 85 to 239 million people as compared to only considering national dynamics (Merkens, et al., 2016). Under the five SSPs and without sea level rise, the population living in the LECZ increases from 640–700 million in 2000 to over one billion in 2050 under all SSPs, and then declines to 500–900 million in 2100 under all SSPs, except for SSP3, for which coastal population reaches 1.1–1.2 billion (Jones and O'Neill, 2016; Merkens, et al., 2016).

A number of new studies have assessed current and future flood risk in terms of the expected number of people affected and monetary average annual losses (AAL) at global levels (Hinkel et al., 2014; Diaz, 2016; Lincke and Hinkel, 2017; Brown et al., 2018; Nicholls et al., 2018), and at the level of major cities around the world (Abadie et al., 2016; Hunter et al., 2017; Abadie, 2018). All of these studies take into account a sea level rise scenarios range wider than the *likely* range of AR5, which is consistent with the range of projections assessed in this report (Section 4.2.3).

Without considering adaptation, there is *high confidence* that sea level rise will have disastrous consequences. For example, considering 21st century sea level rise scenarios of 25–123 cm and uncertainties in elevation data, population data and socio-economic scenarios, Hinkel, et al. (2014) find that 0.2%–4.6% of global population is expected to be flooded annually in 2100, with AAL amounting to 0.3%–9.3% of global GDP. Using the probabilistic city-level scenarios of RCP8.5 from Kopp et al. (2014), Abadie et al. (2016), estimate AAL for European cities, with the biggest losses in 2100 occurring in Istanbul, Odessa, Izmir and Rotterdam (USD5–10 billion; not discounted). Extending this analysis to 120 cities globally, Abadie (2018) find that New Orleans and Guangzhou Guangdong rank highest with AAL above USD1 trillion (not discounted) in each city.

At the same time there is *high confidence* that coastal protection is generally very effective in reducing flood risks during 21st century sea level rise, and also *medium confidence* that it is economically efficient for densely populated areas (Hinkel et al., 2014; Diaz, 2016; Brown et al., 2018; Hinkel, 2018; Lincke, 2018; see also Section 4.4.4.2). For example, Hinkel, et al. (2014) find that coastal protection reduces people flooded and AAL by 2 to 3 orders of magnitude, with global annual investment and maintenance costs of USD12–71 billion in 2100. Coastal protection is widespread today (Section 4.4.2) and there is also *high agreement* that this will continue to be so in the coming decades, as coastal societies have a long history of adapting to coastal environmental change (Kraus, 1996; Charlier, et al., 2005; VanKoningsveld, et al., 2008). For example, some Asian coastal megacities in river deltas have experienced, and adapted to, relative SLR of several meters caused by land subsidence during the 20th century (Kaneko and Toyota, 2011). Hence, sea level rise impacts assessed without adaptation cannot provide an adequate characterization of future coastal flood risks. It is, however, difficult to project how adaptation will play out exactly as this generally entails major social challenges (Section 4.4.6) and is economically less favorable for rural and less densely populated areas and small islands states (Section 4.4.4). In any case, a *likely* impact of SLR will be a diverging world, with richer and densely populated areas behind dikes and poorer less densely populated areas struggling with SLR impacts, and eventually retreating from the coast Hinkel (2018).

Since AR5, the literature has also started to explore a range of other critical dimensions of uncertainty relevant for assessing current and future coastal flood risk. At global scales, uncertainty in socio-economic development, digital elevation methods, emission scenarios, and sea level rise within a given emission scenario, are roughly at equal footing with respect to determining the magnitude of flood risks in the 21st century (Hinkel et al., 2014). At a European level, it was found that the number of people living in the 100-year coastal floodplain can vary between 20% and 70% based on the use of different inundation models and the inclusion or exclusion of wave set up (Vousdoukas, et al., 2016). Using elevation data from local sources instead of global elevation data can result in differences of about 50% in flood damages arraigned (Wolff, et al., 2016). Comparing damage functions attained in different studies for European cities, Prahl et al. (2018) find up to four-fold differences for floods above 3m. Another major sources of uncertainty relates to uncertainties in present-day extreme sea levels due to the application of different extreme value methods (Wahl et al., 2017; Section 4.2.2.5). A comprehensive assessment of uncertainty across these dimensions is missing.

4.3.3.3.2 Projected global impacts of enhanced erosion on human systems

A single global study has assessed the global human and economic impact of coastal erosion (Hinkel, 2013). Without adaptation, about 6000–17,000 km^2 of land is expected to be lost due to enhanced coastal erosion due to SLR during the 21st century, leading to a displacement of 1.6–5.3 million people and associated (not discounted) cumulative costs of USD300–1000 billion. Beach and shore nourishment following annual cost-benefit optimisation including the tourism added value would cost about USD65–220 billion (not discounted) and would reduce 21st century impacts of cumulative land loss by 8–14%, forced migration by 56–68% and the cost of forced migration by 77–84% (not discounted).

4.3.3.3.3 Coastal agriculture

SLR will affect agriculture mainly through land submergence, soil salinization due to marine flooding, salinization and reduction of groundwater lenses, and land loss due to permanent coastal erosion. This translates in effects on production and food security, especially in heavily coastal agriculture-dependent countries such as Bangladesh (Khanom, 2016). Recent literature confirms that it is already a major problem for traditional agriculture in deltas (Wong, 2014) and low-lying island nations where, for example, taro patches are threatened (Nunn et al., 2017b). Taking the case of rice cultivation, recent works emphasize the prevailing role of combined surface elevation and soil salinity, e.g., in the Mekong delta (Smajgl et al., 2015) and in the Ebro delta (Genua-Olmedo et al., 2016), estimating for this latter a decrease in the normalized rice production index from 61.2% in 2010 to 33.8% by 2100 in a 1.8 m SLR scenario. For seven wetland species occurring in coastal freshwater marshes in central Veracruz of the Gulf of Mexico, an increase in salinity was shown to affect the germination process under wetland salt intrusion (Sánchez-García et al., 2017). In coastal Bangladesh, salinity is projected to have an unambiguously negative influence on all dry-season crops over the next 15–45 years (especially in the South-West; Kabir et al., 2018), as well as oilseed, sugarcane and jute cultivation was reported to be already discontinued due to challenges to cope with current salinity levels (Khanom, 2016). Salinity intrusion and salinization can trigger land use changes towards brackish or saline aquaculture such as shrimp or rice-shrimp systems with impacts on environment, livelihoods and income stability (Renaud et al., 2015). However, increasing salinity is only one of the land use change drivers along with e.g. policy changes, and market prices (Renaud et al., 2015).

4.3.3.3.4 Coastal tourism

While SLR will *likely* affect coastal tourist destinations (e.g., beaches), future attractiveness will also depend on changes in air temperature, seasonality and sea surface temperature (including induced effects such as invasive species, e.g., jellyfishes, and disease spreading; Burge et al., 2014; Weatherdon et al., 2016). Future changes in climatic conditions in tourists' areas of origin will also play a role in reshaping tourism flows (Bujosa and Rosselló, 2013; Amelung and Nicholls, 2014), in addition to non-climatic components such as, e.g., accommodation and travel prices, resort's facilities, and tourists' and tourism developers' perceptions of climate-related changes (Shakeela and Becken, 2015). Since AR5, forecasting climate change effects on global-to-local tourism flows remains challenging (Rosselló-Nadal, 2014; Wong, 2014), especially the impacts of SLR *per se*. There are also concerns about the effect of SLR on tourism facilities, for example, in Ghana (Sagoe-Addy and Addo, 2013), all the more that tourism infrastructures contribute to environmental degradation such as coastal erosion, for example (Section 4.3.2). Again, forecasting is constrained by the lack of scientific studies on tourism stakeholders' long-term strategies and adaptive capacity.

4.3.3.3.5 Coastal fisheries and aquaculture

Recent studies support the AR5 conclusion that ocean warming and acidification are considered more influential drivers of changes in fisheries and aquaculture than SLR (Larsen et al., 2014; Nurse et al., 2014; Wong, 2014). The negative effects of SLR on fisheries and aquaculture are indirect (i.e., through adverse impacts on habitats (e.g., coral reef degradation, reduced water quality in deltas and estuarine environments, and soil salinization, etc.), as well as on facilities (e.g., damage to harbours). This makes future projections on SLR implications for coastal and marine fisheries and aquaculture an understudied field of research. Conclusions only state that future impacts will be highly context-specific due to SLR local manifestations as well as to especially local fishery-dependent communities' ability to adapt to alterations in fish and aquaculture conditions and productivity (Hollowed et al., 2013; Weatherdon et al., 2016). Salinity intrusion also contributed to conversion of land or freshwater ponds to brackish or saline aquaculture at many low-lying coastal areas of South-East Asia such as in the Mekong delta in Vietnam (Renaud et al., 2015).

4.3.3.3.6 Social values

Defined as 'the "lived values" of coastal places that are most at risk from sea level rise' (Graham et al., 2013, p. 49), social values offer a wider perspective on impacts on human systems, e.g., complementary to quantitative assessments of health impacts (e.g., loss of source of calories, food insecurity; Keim, 2010). They also offer an opportunity to better consider immaterial dimensions (e.g., some cultural ecosystem services; Fish et al., 2016), as well as context-specificities in valuing both physical/ecological/human impacts' importance for and distribution within a given society. This is a very emerging field of research (no detailed mention in AR5) due to the profoundly transdisciplinary and qualitative nature of the topic. Graham et al. (2013) advance a 5-category framing of social values specifically at risk from SLR: health (i.e., the social determinants of survival such as environmental and housing quality and healthy lifestyles), safety feeling (e.g., financial and job security), belongingness (i.e., attachment to places and people), self-esteem (e.g., social status or pride that can be affected by coastal retreat), and self-actualisation (i.e., people's efforts to define their own identity). In addition, a growing issue relates to territorial sovereignty, from entire nations such as atoll countries (Yamamoto and Esteban, 2016), to parts of countries and individual properties (Marino, 2012; Maldonado et al., 2013; Aerts, 2017).

4.3.4 Conclusion on Coastal Risk

The sections above demonstrate that despite areas of uncertainty on the extent and rate of sea level change (Section 4.2), expected SLR represents a major vehicle for risks on the century scale or sooner (*high* *agreement, medium evidence*). The vast majority of low-lying coasts around the globe, whether in the Global North or South, urban or rural, continental or island, at any latitude, are affected (Cross Chapter Box 5). This chapter also shows that risk will not only result from SLR, but also from SLR interactions with other climate- and ocean-related changes (e.g., extreme events), from the sensitivity of natural coastal systems (Section 4.3.2.1) and from anthropogenic processes such as coastal urbanization, population growth and changes in lifestyles (Section 4.3.2.2). This section highlights such a complementary issue (compound event) as well as first insights on a synthesis of risk induced by SLR at the global scale.

4.3.4.1 Compound Events

A compound event occurs when impacts of a climate event or trend interact with or precondition the impacts of a simultaneous or subsequent event (SREX Sections 1.2.3.2 and 3.1.3). One statistical definition is 'an extreme impact that depends on multiple statistically dependent variables or events' (Leonard et al., 2013). For example, tropical cyclones following similar paths can lead to compound damages with greater than additive consequences. Extreme impacts from compounding can occur even when one or more of the individual events is not extreme or if the events are of different types, such as simultaneous coastal and riverine flooding (SREX Section 3.1.3). Impacts of events that seem to be spatially and temporally distinct can also interact to increase losses (Hillier et al., 2015). An initial event may lead indirectly to compound effects by reducing capacity to respond to subsequent events, for example, by exhausting financial and human resources available for disaster response (see Section 6.6 of this report).

AR5 definitions refer to "compound risk" and point to geographic areas where compound risk is particularly relevant to SROCC: "examples include the Arctic (where thawing and sea ice loss disrupt land transportation, buildings, other infrastructure, and are projected to disrupt indigenous culture); and the environs of Micronesia, Mariana Island, and Papua New Guinea (where coral reefs are highly threatened due to exposure to concomitant sea surface temperature rise and ocean acidification)" (Oppenheimer et al., 2014, p. 1042). Cities situated on river deltas are especially subject to high levels of compound risk (Oppenheimer et al., 2014).

Research on the characteristics of compound events is limited. Among the compound hazards analyzed are episodes of extreme heat and humidity (Fischer and Knutti, 2013), tornado outbursts in the US (Tippett et al., 2016), drought and extreme heat (AghaKouchak et al., 2014) in California, storm tides arising from the combination of sea level rise and tropical cyclones (Little et al., 2015), extreme storm surge in combination with extreme precipitation in the US (Wahl et al., 2015; Moftakhari et al., 2017), and projected increased clustering of extremely hot days (JW Baldwin, 2018). There is *high confidence* that the frequency of compound events related to extreme marine heat (such as those damaging to Arctic and coral reef systems), and the associated risk, will increase over the 21st century.

We find no studies analyzing impacts and adaptation to compound events related specifically to SLR. Examples from other sectors may be instructive. For example, SREX provided a detailed case study of Mongolia's Dzud events that are characterized by large losses of livestock and encompass episodes of extreme cold, windstorms, drought, and/or heavy snowfall. Consequences have included undermining of livelihoods, migration, and urbanization Murray et al. (2012). Similarly, teleconnected flooding and drought linked to El Nino, resulting in crop yield declines and severe food security disruptions worldwide, as demonstrated by a hypothetical scenario (Lunt et al., 2016), provides an example of climate events at widely separated locations resulting in compound impacts.

4.3.4.2 Reasons for Concern

Low-lying islands, coasts and communities provide relevant illustrations of some of the Reasons for Concern (RFCs) developed by the IPCC since AR3 (McCarthy, 2001) and describing potentially dangerous anthropogenic interference with the climate system (in reference to one of the core objectives of the UNFCCC). In particular, the RFCs illustrate the risks to unique and threatened systems (RFC1), and risks associated with extreme weather events (RFC2) and with the uneven distribution of impacts (RFC3). The AR5 Synthesis Report (IPCC, 2014) developed two additional RFCs that are pertinent to ocean and coastal environments: one is based on risks to marine species arising from ocean acidification and another on risks to human and natural systems from sea level rise. Recent scientific advances lead to a re-evaluation of all

RFCs (O'Neill et al., 2017). Despite the difficulty in attributing observed impacts to SLR *per se* (Section 4.3.3.1.4), O'Neill et al. (2017) estimate that risks related to SLR are already detectable and would increase rapidly, so that 'high risk may occur before the 1 m level (above the 1986–2005 level; 1 m is a benchmark for the sea level rise RFC) is reached'. In addition, limits to coastal protection and ecosystem-based adaptation exist above 1m rise by 2100. In SROCC, estimated SLR is revised upward due to recent studies of Antarctica's potential contribution (Section 4.2.3).

[PLACEHOLDER FOR SECOND ORDER DRAFT: RFC revisions to be added once sea level projections become available and impacts are assessed accordingly].

As an indicator of risk, the sea level rise RFC has several limitations. It does not fully integrate either the possibility of abrupt changes (6.2), the crossing of environmental and/or anthropogenic tipping points (6.2) or changes in extreme sea levels (Sections 4.2.2.5 and 4.2.3.3). [PLACEHOLDER FOR SECOND ORDER DRAFT: to account for these effects may be possible, depending on results from other chapters]. Furthermore, the RFCs and associated burning embers were developed at a global scale (see Oppenheimer et al. (2014), Gattuso et al. (2015), Magnan et al. (2016b), and O'Neill et al. (2017)). As a result, in its current form, the framework does not describe the geographical variability of risk from one low-lying coastal area to another, except with respect to RFC3.

4.4 Responses to Sea Level Rise

4.4.1 Introduction

This chapter describes the variety of responses available, where and how they have been applied, their costs, benefits and co-benefits, frameworks for appraising and choosing appropriate options, as well as limits and barriers of their implementation.

4.4.2 Types of Response Measures

Following earlier IPCC Reports (Nicholls, et al., 2007b; Wong, 2014), we distinguished between four fundamentally different types of responses to sea level rise and its impacts (Table 4.6):

Protection measures reduce the chances of, or completely prevent, coastal impacts from occurring. These include three sub-categories of measures. First, there are hard engineering structures such as dikes, seawalls, breakwaters and surge barriers to protect against flooding and erosion, or barriers and barrages to also protect against salt water intrusion (Nicholls, et al., 2018). Second there are sediment-based measures such as beach and shore nourishment, dunes (also referred to as soft structures) and land raising. Third, there are ecosystem-based measures (EBM) that use ecological features such as reefs and coastal vegetation to provide adaptation benefits. EBM protect the coastline in three ways: a) by attenuating the energy, and hence height, of incoming waves and in some cases, storm surges; b) by trapping and stabilizing coastal sediments and reducing rates of erosion; c) by raising shoreline elevations through the build-up of organic matter and detritus (Shepard, et al., 2011; McIvor, et al., 2012a; McIvor, et al., 2012b; Cheong, 2013; McIvor, et al., 2013; Spalding, et al., 2014). Hybrid approaches combining elements of hard, sediment-based and ecosystem-based measures are also common (Sutton-Grier et al., 2015; Small-Lorenz, 2016). Protection is generally not a response to sea level rise only, but also to socio-economic development (e.g., raise dikes with increasing affluence) as well as current coastal flood and related hazards.

Advance measures create new land by building seaward and upwards. This includes large-scale land reclamation above sea levels by land filling with pumped sand or other fill material, planting vegetation in order to support natural accretion of land and surrounding low areas with dikes, termed polderisation. Advance has a long history in most areas where there are dense human populations and a shortage of land such as around the southern North Sea (Germany, the Netherlands, Belgium and England) and China (Wang et al., 2014a). Hence, in the past it was not primarily a response to SLR, but to a range of drivers including land scarcity and population pressure, as well as management of extreme events. Looking to the future, these advanced land areas will require adaptation, and future advance measures will become more integrated with adaptation and might even be seen as an opportunity in some cases (Linham and Nicholls, 2010; RIBA Royal Institute of British Architects, 2010; Nicholls, 2018a).

Accommodation measures do not prevent coastal impacts from occurring, but reduce vulnerability to these. This covers measures going from addressing physical issues (e.g., standards to raise buildings floor level in small islands) to the diversification of livelihoods (e.g., shift in harvesting activities in the Arctic, cultivation of saline-tolerant crops, landscape restoration for tourism purposes), to institutional approaches (e.g., community participation in local government decision-making, establishment of marine parks and protected areas, integrated coastal management plans) (Nurse, et al., 2014; Wong, 2014). Physical accommodation is accomplished by building regulation and codes which apply standards for new construction and retrofitting existing properties. Accommodation measures for salinity intrusion include salt tolerant crop varieties and changing land use from, e.g., freshwater rice paddy to brackish/salt shrimp aquaculture (Nicholls, 2018b).

Retreat measures reduce exposure to coastal impacts by moving people, infrastructures and activities out of the coastal hazard zone or steering future development away from it. Retreat includes migration, permanent or semi-permanent move by a person, at least for one year (Adger, et al., 2014), forced displacement and planned relocation (also called planned retreat or managed realignment), which is typically initiated, supervised and implemented at the State-to-local level, and develops form small communities and individual assets to, more rarely, large populations (Wong, 2014; Hino, et al., 2017).

Table 4.6: Overview of exemplary coastal response measures and the impact they address.

Categories		Measures		Impact Addressed						
				Flooding	*Submergence*	*Erosion*	*Impeded Drainage*	*Salinity Rivers*	*Salinity Aquifers*	*Wetland Change & Loss*
Protect	*Hard*	Sea wall		X	X					
		Sea dike		X	X					
		Breakwater		X		X				
		Groynes				X				
		Fixed barrage/closure dam		X	X			X		
		Storm surge barrier		X						
		Saltwater intrusion barriers						X	X	
	Sediment-based	Land raising	Through artificial filling	X	X	X	X		X	
			Through controlled natural sedimentation (e.g., tidal river management)	X		X			X	
		Shore and beach nourishment	Emergency nourishment	X		X				

			Periodic nourishment	X		X				
			Mega-nourishment (e.g., Sand engine)	X		X				
		Dunes	Conservation	X		X				
			Restoration	X		X				
	Ecosystem-based	Vegetation (marshes/mangroves, seagrasses/kelp, mussel beds)	Conservation	X		X				X
			Restoration	X	X	X				X
		Reefs (Coral/ Oyster)	Conservation	X		X				
			Restoration	X		X				
Advance	*Land reclamation*	Through land filling with pumped sand or other fill material		X	X	X	X			
		Through sediment accretion by vegetation and natural processes		X	X	X	X			X
		Through polders (Enclosed areas with dikes and improved drainage)		X	X	X	X			
Accomodation	*Physical*	Floor raising		X	X					
		Flood-proofing of buildings		X						
		Drainage systems					X	X	X	
		Land-use change (salt tolerant crops, aquaculture, etc.)		X	X					
	Institutional	Early warning system		X						
		Insurance systems		X		X				
		Emergency planning		X						
Retreat	*Unplanned*	Migration		X	X	X	X	X	X	X
		Displacement		X	X	X	X	X	X	X
	Planned	Managed realignment		X	X	X	X	X	X	X
		Setback zones		X	X	X	X	X	X	X

4.4.3 Observed Responses Across Geographies

4.4.3.1 Observed Changes in Coastal Policies and Planning

[PLACEHOLDER FOR SECOND ORDER DRAFT]

4.4.3.2 Hard Protection

Hard and sediment-based measures are widespread around the world e.g. in the Pacific region (Paeniu et al., 2015), Northern Ireland (Cooper et al., 2016) and New York City (Rosenzweig and Solecki, 2014), although it is difficult to provide estimates on how many people are protected by them. Currently, at least 20 million people living up to several meters below normal high tides are protected by hard structures in countries such as Belgium, Canada, China, Germany, Italy, Japan, the Netherlands, Poland, Thailand, the UK, and the USA (Nicholls, 2010), but many more people living above high tides are also protected through hard structures in major cities around the world. Gittman et al. (2015) estimate, for example, that 14% of the total US coastline has been armored. In coastal lowlands land reclamation through polders has been extensively used, or the land has been filled with sediment above normal tidal levels. On some steep coasts where there is little flat land, such as Hong Kong, higher areas have been lowered to create fill material to build land out into the sea. Land claim has taken place in all major coastal cities to some degree, even if just for the creation of the port and harbour area. Globally, it is estimated that about 33,700 km^2 of land has been gained from the sea during the last 30 years (about 50% more than has been lost), with the biggest gains being due to land reclamation in places like Dubai, Singapore and China (W. Wang, et al., 2014; Donchyts, 2016). In Shanghai alone, 590 km^2 land has been reclaimed during the same period (Sengupta et al., 2018) and significant further land claim is expected in land scarce situations such as China, Japan and Singapore.

Contemporary engineered coastal defences have been mostly developed in response to storm tides, wave-run-up, and the pounding from waves, and rarely in anticipation to future climate change impacts including SLR (e.g., Oppenheimer and Alley (2016)). If sea level rise is considered in the planning process, than it typically assumes an increase of the height of coastal defences by an amount equivalent to the regionally projected SLR height, although SLR driven changes in wave and tides characteristics amplify the expected design heights of the infrastructure by an average of 48–56%, when compared with design changes caused by SLR alone (Arns et al., 2017). There are new approaches in coastal protection, which account more for dynamic adjustments over time and are based on gradually implemented no- or low-regret solutions to adjust to the latest available knowledge about climate change and its impacts on the coast such as London´s TE2100 Plan (Environmental Agency, 2012; World Ocean Review, 2017).

While recognizing the potential benefits to be expected from technical and engineering options in terms of risk reduction for human settlements, the AR5 remains cautious by stating that such a conclusion is supported by limited evidence and high agreement in low-lying coastal areas in general (Wong, 2014), and by medium evidence and medium agreement in small islands specifically (Nurse et al., 2014). The abundant recent literature reports growing concerns about limitations and sometimes medium-term counterproductive effects of such measures, e.g., on ecosystems (Gittman et al., 2016). In contexts lacking adequate funds (and mechanisms guaranteeing their sustainability), policies (especially maintenance programs, building codes and strategic planning) and technical skills (Nurse et al., 2014), seawalls can be illustrative of a maladaptive option, as such measures can result in development intensification behind such protection that increases risk in the face of relentless SLR. Levees in New Orleans are illustrative examples (Burby, 2006; Kates et al., 2006) of coastal areas with considerable investments into engineered protection (Kates et al., 2006; Rosenzweig and Solecki, 2014; Cooper et al., 2016).

In addition, the recent literature confirms that decisions about the ‘best responses’, either to react or anticipate, remains challenging. For example, in the Outer Hebrides, UK, ‘community driven decision-making process was shown to lead to the erection of a coastal defence bund against the advice of coastal experts’ (Young et al., 2014). Contrasting views also emerge among stakeholders (Evans et al., 2017) as well as different populations groups, e.g., in Japan about preferences for coastal constructions vs. ecosystem-based approaches (Imamura et al., 2016), or in South-East England about the willingness to pay for coastal defences (Jones et al., 2015).

These last points confirm a major conclusion of the AR5-cycle, for instance, the relevance of technical and engineering options [as well as others] to enhance long-term adaptation is critically context-specific, depending on both natural and human circumstances (*high evidence*, *high agreement*). Growing literature also advocates in favour of combinations of options and their sequencing through time (see Section 4.3.4.2). An underlying issue about the choice for technical and engineering options remains, as for any other options, the difficulty to integrate a long-term challenge such as SLR.

4.4.3.3 Ecosystem-based Measures (EbA) to Sea Level Rise and Related Coastal Hazards

Relative to hard adaptation measures whose global distribution is not known in detail (Scussolini, et al., 2015), the current distribution of coastal ecosystems is well-studied, while potential restoration extents are not as well-understood. Ecosystem-based adaptation, by definition, can only exist and function in locations where these ecosystems occur naturally. Habitats like mangroves, salt marshes and reefs probably cover ~40% to 50% (800,000 to 1,000,000 km) of the world's coastlines (Wessel and Smith, 1996; Burke, 2011; Giri, 2011; Mcowen, et al., 2017). However, there is no clear estimate on the global length of coastline covered by habitats. The spatial resolutions of these estimates are often higher than most global coastline length estimates to capture individual habitat extents. Mangroves occur on tropical and subtropical coasts, totaling around 13,776,000 ha across 118 countries (Giri, 2011). At least 150,000 km of coastline in over 100 countries benefit from the presence of coral reefs (Burke, 2011). Extents of other coastal habitats are less well known: salt marshes are estimated to occur in 99 countries, with nearly 5,500,000 ha of these mapped across 43 countries (Mcowen, et al., 2017). These estimates allow calculation of the economic values of each ecosystem to people and property at multiple spatial scales. For example, globally, coral reefs are estimated to protect over 100 million people from wave-induced flooding (Ferrario, 2014). Ecosystem-based adaptation (EbA) or more generally nature based solutions (NbS) are key instruments towards a transition to sustainability and it is a political, scientific and technological challenge and endeavor (Scarano, 2017). EbA and NbS is increasingly discussed and implemented (Cohen-Shacham et al., 2016; Wamsler et al., 2016) as one element of the overall adaptation strategy to help people to adapt to the adverse effects of climate change. While the term Ecosystem-based adaptation (EbA) is used in this section, related solutions and implementation examples in the literature might fall under the categories of Ecosystem-based disaster Risk Reduction (Eco-DRR), nature-based solutions (NbS), Green Infrastructure (GI), 'ecological engineering', 'Building with Nature' or hybrid infrastructure depending on their specific aim or when designed and implemented. Main challenges identified in the IPCC AR 5 were the low number of implemented ecosystem-based solutions to assess either the risks or the benefits comprehensively (AR5 Cross Chapter Box on EbA). Today, although engineered and technological adaptation options are still the most common responses at the coast, there is growing number of implemented ecosystem-based and hybrid solutions worldwide. Specifically in the coastal areas, a number of countries and communities implementing ecosystem-based coastal adaptation measures and these measures are increasingly mainstreamed into national plans, strategies and targets, including national adaptation programmes of action (NAPAs) under the United Nations Framework Convention on Climate Change (UNFCCC), national biodiversity strategies and action plans (NBSAPs) under the Convention on Biological Diversity (CBD), disaster management plans, and development policies. A review of examples of mainstreaming ecosystem-based solutions through these national plans, strategies and targets has been synthesized in CBD Tech. Series No. 85 (Lo, 2016). Other examples include the Adaptation Fund (AF), which supports a number of EbA projects such as using EbA approaches in the Seychalles along the shorelines of the Granitic Islands to reduce the risks of climate change induced coastal flooding and to mainstream EbA into development planning and financing. The AF also finances a project in India aiming to overcome the consequences of sea level rise and seawater inundation through restoration of degraded mangroves and demonstration of Integrated Mangrove Fishery Farming System. Achieving synergies in implementation of EbA, with for example disaster risk reduction, mitigation or biodiversity protection is increasingly recognized as a mainstreaming option. Examples include the inclusion of ecosystem-based measures for mitigation and adaptation in National Biodiversity Strategies and Action Plans (e.g., Uganda, Cameroon, Sri Lanka), a strategies on forests that includes ecosystem-based adaptation and mitigation actions as well as references to the National Biodiversity Strategy (e.g., Peru), or ecosystem-based measures that seek synergies among mitigation and adaptation in the Nationally Determined Contributions to UNFCCC (e.g., Mongolia, China; Epple, 2016). Table 4.7 provides a selection of open sources where EbA examples are presented worldwide.

Table 4.7: Databases of EbA and other nature-based measures incl. coastal applications

Scope and geography	Description
Database on ecosystem-based approaches to Adaptation (UNFCCC) *Global*	An initiative under the Nairobi work programme to provide examples of ecosystem-based approaches to adaptation, supplementing information to FCCC/SBSTA/ 2011/INF.8, mandated by the SBSTA at its thirty-fourth session under the Nairobi work programme. *Link:* http://www4.unfccc.int/sites/NWP/Pages/soe.aspx
Adaptation Fund Projects & Programmes Coastal Zone Management	The database provides an overview about approved projects including their aim, implementation status and implementing agency. Link: https://www.adaptation-fund.org/projects-programmes/project-sectors/coastal-zone-management/
International Climate Initiative (IKI) Projects Adaptation	The database provides an overview about funded projects by IKI including their aim, implementation status and implementing agency. Link: https://www.international-climate-initiative.com/en/projects/
Climate Change Adaptation Database - Integrating Biodiversity into Climate Change Adaptation Planning (CBD) *Global*	The database provides web-based guidance on the integration of biodiversity within adaptation planning. It gathers information tools and case studies from a number of relevant partners. It provides links to scientific studies and other resources on biodiversity-related climate change adaptation. These examples can assist managers and governments to find adaptation options that will not have a negative impact on biodiversity. Link: https://adaptation.cbd.int/options.shtml#sec1
PANORAMA – Solutions for a healthy planet (GIZ, IUCN, UN Environment, GRID Arendal, Rare) *Global*	An interactive platform and database of specific, applied examples of successful NBS, EbA and Eco-DRR processes or approaches structured according to regions, ecosystems, specific thematic areas, governance and hazards addressed. Useful for identifying different targets (Aichi, Sendai Framework, SDGs, NDC) and outlining challenges. *Link:* http://panorama.solutions/en/explorer/grid/1042
Natural Water Retention Measures catalogue (EU) *Europe*	NWRM cover a wide range of actions and land use types. Many different measures can act as NWRM, by encouraging the retention of water within a catchment and, through that, enhancing the natural functioning of the catchment. The catalogue of measures hereunder is sorted by sector. It has been developed in the NWRM project, represents a comprehensive but non prescriptive wide range of measures. *Link:* http://nwrm.eu/measures-catalogue
Naturally resilient communities (US National Planning Association) *North America*	This database allows to explore over 50 solutions and case studies on nature-based solutions and included case studies of successful projects from across the US to help communities learn more and identify which nature-based solutions might work for them. The explorer allows to filter by cost, region, hazards, and more. *Link:* http://nrcsolutions.org/ http://nrcsolutions.org/

Equator Initiative, Solutions Database *Global*	The Solutions Database helps to learn how local communities and indigenous peoples around the world are making possible the achievement of the UN Sustainable Development Goals through nature-based actions. Use the interactive map or the filter to find coastal solutions Link: https://www.equatorinitiative.org/knowledge-center/nature-based-solutions-database/

Many ecosystem-based practices involve community participation and ownership, and thus create synergies with Community-based adaptation (CBA)—a process that is led by communities, based on their priorities, needs and capacities (Reid, 2016). While differing in theory, often both approaches are used in local adaptation efforts and are indistinguishable in the field (Reid, 2016). There is a large number of local ecosystem-based actions on the ground to retain and or restore coastal ecosystem functions, including measures that will provide natural protective measures in the face of SLR, in projects in global north and south ranging from mangrove plantings to dune rehabilitation, etc. which are not necessarily reported under any of the global frameworks but contribute to adaptation to SLR.

4.4.3.4 Retreat Responses including Human Mobility

Climate change is exacerbating millions of people's vulnerabilities by increasing pressure on resources and land, with expected consequences on environmental-induced human mobility (i.e., mobility, displacement, relocation). *(i)* Mobility has been defined in the AR5 as the permanent or semi-permanent move by a person, at least for one year and involving crossing an administrative, but not necessarily a national, border (Adger et al., 2014). Changes in mobility patterns can be responses to both extreme weather events and longer-term climate variability and change. Mobility due to environmental change is not a new phenomenon, for example in the Pacific (Connell, 2012; Janif et al., 2016), and it is growingly considered as a potentially effective adaptation strategy in response to SLR (Shayegh et al., 2016; Hauer, 2017; Morrison, 2017). *(ii)* Displacement refers to forced forms of human mobility, and climate change is projected to increase displacement trends over the 21stcentury (*medium evidence, high agreement*; Shayegh, 2017). Significantly higher risk are found in lower-income countries, small islands and low-lying coastal areas (e.g. in Vietnam, Bangladesh, Egypt, Malaysia, Thailand, Myanmar, the Philippines, Indonesia, China and Iraq), and especially among people lacking the resources for planned mobility (Milan and Ruano, 2014; Logan et al., 2016). This is also true in developed countries, as reported by (Logan et al., 2016; p. 1511) after hurricanes in the Gulf Coasts, USA: 'advantaged population groups are more *likely* to move out of or avoid moving into harm's way while socially vulnerable groups have fewer choices.' The poorest households are indeed significantly more *likely* to endure material and human losses following a natural hazard, and repeated losses of livelihood make them more vulnerable to future risk. Economic and human losses are also at risk from salinization, tidal surge, erosion, and household location, making the achievement of sustainable development in low-elevation deltas, e.g., a major challenge for the coming decades (Hajra et al., 2017). It is estimated that SLR in a 2°C warmer world could submerge land currently home to 280 million people globally by the end of this century (Strauss et al., 2015), which raises major concerns according to the current effects of weather-related disasters on human displacement (i.e. 17.5 and 19.2 million people displaced in 2014 and 2015, respectively; IDMC, 2016). It however remains scientifically challenging to estimate future displacement associated with SLR, recent literature emphasizing that economic and political factors still are powerful drivers of human mobility associated with disasters (Stapleton et al., 2017). Despite this, international mobility and displacement gained major attention over the last decade in both scientific arena (UNHCR, 2016b; UNHCR, 2016a) and the international policy community. Recently, the Paris Agreement instituted a dedicated taskforce, and The Nansen Initiative has been initiated by 110 countries to address the serious legal gap currently existing about cross-border human movements (https://www.nanseninitiative.org/secretariat/). Evidence of mobility and displacement however remain debated. In small islands for example, the AR5 specifies that 'evidence of human mobility as a response to climate change is scarce [and] there is no evidence of any government policy that allow for climate "refugees" from islands to be accepted into another country' (Nurse et al., 2014; p. 1625). Environmental Justice Foundation (Environmental Justice Foundation, 2017) suggests that sea level rise will lead to the displacement of hundreds of millions of people by 2100. On the other hand, experience of mobility as a social response to climate change risks has been documented in the Caribbean (Rivera-Collazo et al., 2015), and the government of Vietnam promotes rural populations' mobility to industrial areas with labour needs as a response to future climate change (Collins et al., 2017). *(iii)* Relocation – also called coastal retreat or planned resettlement—offers a third option for human mobility in the face of SLR (Wong, 2014; Hino et al., 2017). It is typically initiated, supervised and implemented at the State-to-local level, and develops form small communities and individual assets to, more rarely, large populations (Hino et al., 2017). While usually discussed after an extreme event, such as Xynthia storm in France (Genovese and Przyluski, 2013) or Hurricane Sandy in the USA (Bukvic and Owen, 2017), relocation plans generally target the reduction of long-term environmental risks including SLR (McAdam and Ferris, 2015; Morrison, 2017). In many circumstances, resettlement leads to decline of well-being and livelihood for those resettled, at least when communities are dispersed (Kura et al., 2017), and generally raised controversial views (Genovese and Przyluski, 2013; Ford et al., 2015; Nordstrom et al., 2015; Bukvic and Owen, 2017; Hino et al., 2017; Jamero et al., 2017).

4.4.4 Economic Costs, Benefits, and Co-benefits of Response Measures

Since AR5, more information on economic costs and benefits of response measures has become available. This information is assessed here in terms of capital cost, maintenance costs, intended biophysical and monetary benefits, as well as co-benefits (positive) and drawbacks (negative) that arise next to the intended benefits. Finally, we synthesis this information across types of measures into general design considerations.

4.4.4.1 Hard and Sediment-based Protection Measures

4.4.4.1.1 Costs of hard measures

As shown in Table 4.8, protection cost can be expressed as cost per unit length protected and increase in height of the structure. This was recognised by Dronkers (1990) who developed a number of unit costs of defences, feeding into the first estimate of the global cost of adapting to a 1 m rise in sea in the IPCC First Assessment Report. These costs have been taken forward in a series of global assessments as reviewed by Jonkman (2013) and continued to feed into global adaptation cost estimates (Hinkel et al., 2014; Nicholls, 2018c). The variance in costs has been examined by a few authors such as Jonkman (2013) but in general there has been limited systematic data collection across sites, although useful national guidance does exist in some cases (Environment Agency, 2015). Defences depend on good maintenance to remain effective and annual maintenance budget of 1% to 2% of capital costs can be expected for this purpose (Jonkman, 2013). For some types of infrastructure such as surge barriers, costs could be higher, but these costs are poorly described and hence uncertain (Nicholls, et al., 2007a). Adaptation to saltwater intrusion is more complex and bespoke than adaptation to flooding and erosion, and there is less experience to draw upon.

Table 4.8: Capital and maintenance costs of hard protection measures.

Measure	Capital Cost (in million USD per km of coastline and 10 cm SLR/height unless stated otherwise)	Annual Maintenance Cost (% of capital cost)
Sea Wall	0.04–2.75 (Linham, et al., 2010)	1 to 2% per annum (Jonkman, 2013)
Sea Dike	0.09–2.92 (Jonkman, 2013)	1 to 2% per annum (Jonkman, 2013)
Breakwater	0.25–1.0 (Narayan, et al., 2016)	1% per annum (Jonkman, 2013)
Storm Surge Barrier	0.5–27 (Jonkman, 2013) or 2.2 (Mooyaart and Jonkman, 2017) million Euro per meter width	1% per annum (Mooyaart and Jonkman, 2017) or 5 to 10% per annum (Nicholls, et al., 2007a)
Saltwater Intrusion Barriers	Limited knowledge	Limited knowledge

4.4.4.1.2 Costs of sediment-based measures

Sediment-based measures are generally costed as the unit cost of sand (or gravel) delivery versus the volumetric demand for beach nourishment. Unit costs range from USD3–15 per m³ sand (Linham, et al., 2010), with some high outlier costs in the UK and New Zealand. Costs are small where sources of sand are plentiful and close to the sites of demand and where the shoreface nourishment delivers the sand to the beach. This situation is found in the Netherlands where the entire open coast is maintained with large-scale shore nourishment (Mulder, et al., 2011) and the innovative sand engine has been implemented as a full-scale decadal experiment (Stive, et al., 2013). The difference between hard and sediment-based measures is that the later are sacrificial and require regular renourishment to maintain the design standard. Hence the costs of nourishment need to be described on a whole-life basis (with appropriate discounting) to reflect the repeated renourishment volumes, but the maintenance costs are generally lower compared to hard engineering. One essential maintenance cost component is regular beach/shoreface monitoring to assess the beach volume. Beach nourishment also often includes a dune which protects against erosion and flooding. While this is a small part of the overall beach volume, it is critical to the defence provided by nourishment, and particular attention needs to be focussed on monitoring and maintaining the dune. Capital costs for dunes will be similar to beach nourishment, although placement and planting vegetation may raise costs. Maintenance costs vary from almost nothing to several million dollars per kilometre, although costs are usually at the lower end of this range (Environment Agency, 2015).

4.4.4.1.3 Benefits, co-benefits and drawbacks

There is *high confidence* that well designed and maintained hard and soft protection provides predictable levels of safety. Hard defenses alter hydrodynamic and morphodynamic patterns and by doing so may induce flooding and erosion problems elsewhere. There is also the risk of locking into hard protection strategies, because protection attracts further economic development in the flood zone, which again lead to further raising defenses. Hard protection hinders or prohibits the onshore migration of geomorphic features and ecosystems causing coastal squeeze (Pontee, 2013). Many hard defences exist in combination with ecosystems such as marshes and mangroves that provide additional protection. If the latter are degraded through coastal squeeze or otherwise, then the defences need to be upgraded, or lower level of safety accepted. The loss of habitats violates many statutes such as the EU Habitats Directive. Softer protection such as beach nourishment preserves beach and associated environments. A significant emerging issue, however, is sourcing the increasing volumes of sand required to sustain beach volumes that are adequate to provide a sustainable level of protection in the face of sea level rise (Roelvink, 2015).

4.4.4.2 Ecosystem-based Protection

4.4.4.2.1 Costs

There is *limited evidence* and *low agreement* on the costs of ecosystem-based measures. The total cost of an ecosystem-based measure depends on several components, including capital costs, maintenance costs, the cost of land and, in some situations, permitting costs (Bilkovic, 2017). Not enough is known about the factors that influence these to be able to generalize or estimate unit costs across large spatial scales. The costs of restoring and maintaining coastal habitats depend on coastal setting, habitat type and project conditions. In general, restoration is cheaper in developing countries and unit restoration costs are lowest for mangroves, higher for salt marshes and oyster reefs and highest for seagrass beds and coral reefs (Table 4.9). Unit restoration costs are typically measured per hectare and vary from less than USD10,000 per hectare for mangroves to more than USD150,000 per hectare for coral reefs (Bayraktarov, 2016; Narayan, et al., 2016). The conservation of coral reefs and other coastal habitats may also entail substantial opportunity costs that are often overlooked (Stewart, et al., 2003; Balmford, 2004; Adams, 2011; Hunt, 2013). Ecosystem-based measures also require periodic maintenance to preserve their coastal protection benefits though there is *limited evidence* for these costs. Maintenance is particularly important in the immediate aftermath of storms, when wetlands and reefs can be damaged by high winds, waves and surges, and further affected by sediments and debris (Smith III, et al., 2009; Puotinen, et al., 2016). Yet, a body of case-studies and an increasing number of targeted public and private financial mechanisms are emerging, that can promote and incentivise the implementation of ecosystem-based measures for adaptation and risk reduction (Colgan, 2017; Sutton-Grier, et al., 2018).

Table 4.9: Costs of ecosystem-based protection

	Capital Costs	Maintenance Costs
Dune Conservation	None	No data available
Dune Restoration	USD3–15 per m^3 sand (Linham and Nicholls, 2010)	No data available

Vegetation Conservation	None	Thinning, clearing debris after storms, etc.: Mangrove: USD5000/ha yr^{-1} in Florida (Lewis, 2001).
Vegetation Restoration (Marshes/Mangroves, Maritime Forests)	Mangroves: USD9,000/ha (median) (Bayraktarov, 2016); USD2,000–13,000/ha in American Samoa (Gilman and Ellison, 2007); Salt Marshes: USD67,000/ha (Bayraktarov, 2016)	No data available
Reef Conservation (Coral/ Oyster)	None	No data available; low
Reef Restoration (Coral/ Oyster)	USD165,600/ha (median) (Bayraktarov, 2016); Oyster Reefs: USD66,800/ha (median) (Bayraktarov, 2016)	No data available; low

4.4.4.2.2 *Physical benefits*

There is *high evidence and high agreement* for the physical effectiveness of ecosystem-based measures for reducing wave heights and storm surges (Doswald, 2012; Lo, 2016; Renaud, et al., 2016). Dozens of independent field and experimental studies have observed and measured the physical benefits provided by natural habitats such as marsh and mangrove wetlands, (Barbier, 2014; Möller, et al., 2014), coral reefs (Ferrario, 2014), oyster reefs (Scyphers, et al., 2011), beaches, dunes, and barrier islands (Stive, et al., 2013; Hanley, 2014) and even submerged seagrass beds (Infantes, 2012). Generally, studies on ecosystem-based measures do not consider beaches and dunes though these habitats are well-established in coastal engineering practice and commonly used as adaptation measures (Hinkel, 2013; Hanley, 2014; Pontee, et al., 2016). A synthesis of 69 field studies of wave attenuation within coastal habitats showed average attenuation rates of more than 30% in mangroves, kelp beds and seagrass beds and nearly 70% in coral reefs and salt marshes (Narayan, et al., 2016). The synthesis also found that coral reefs and salt marshes tend to occur in higher wave energy environments relative to mangroves and seagrass beds. Studies based on field observations, experiments and numerical models elucidate some of the parameters that influence this effectiveness, such as structural complexity in coral reefs (Harris, 2018), vegetation density, height and structural complexity in salt marshes (Möller, et al., 2014) and mangrove forests (Maza, et al., 2016). Salt marsh and mangrove wetlands can also reduce storm surge levels during extreme events (Krauss, 2009; Zhang, et al., 2012; Vuik, 2015). Rates of surge attenuation can vary between 5 and 70 cm/ km (Krauss, 2009; Vuik, 2015) and depend on several storm, wetland and landscape characteristics (Loder, et al., 2009; Wamsley, et al., 2010). In general, ecosystem-based measures, where they are suitable, are recognized as being one important component within a typical suite of risk reduction and adaptation solutions (Spalding, et al., 2014).

4.4.4.2.3 *Economic benefits*

There is *medium evidence* that ecosystem-based measures bring substantial economic benefits, but *low agreement* regarding the actual size of the benefits. The most common value-estimation methods for ecosystem-based measures for risk reduction are the replacement cost approach (Barbier, 2007) and the avoided damages approach (Beck, 2016). Using these methods, studies have shown that coastal habitats can a) save lives during cyclones (Das, 2009); b) reduce millions of dollars in flood damages from storm surges (Barbier et al., 2013; Narayan et al., 2017); c) reduce the required crest heights and maintenance costs for seawalls and dykes (Möller et al., 2001; IFRC International Federation of Red Cross and Red Crescent Societies, 2011; Van Slobbe et al., 2013); and; d) be restored to provide protection from waves at costs 2 to 5 times lower than breakwaters (Ferrario, 2014; Narayan, et al., 2016). For example, the number of people, and total value of residential property that are most exposed to coastal hazards can be reduced by half if existing coastal habitats remain fully intact based on different SLR scenarios in the US (Arkema et al., 2013). However, the benefits of ecosystem-based measures, unlike typical engineering structures, exhibit high natural variability in time and space and depend on multiple physical and biological parameters (Koch, 2009; Pinsky, et al., 2013). This makes it difficult to extrapolate values of physical or economic benefits across geographies.

4.4.4.2.4 *Co-benefits*

There is *high evidence and high agreement* that ecosystem-based measures provide multiple additional co-benefits such as sequestering carbon (Siikamäki, et al., 2012; Hamilton, 2018), facilitating income from tourism (Carr, 2003; Spalding, et al., 2017), enhancing fishery productivity (Carrasquilla-Henao, 2017; Taylor, et al., 2018), improving water quality (Coen, 2007; Lamb, 2017), providing raw material for food, medicine, fuel and construction (Hussain, 2010; Uddin, et al., 2013), and providing a range of intangible and cultural benefits generally difficult to express in monetary terms (Scyphers, et al., 2015). The value of mangroves was shown to be dominated by their coastal protection services (incl. climate mitigation, erosion control and defense against extreme weather events) as opposed to their provisioning services (Emerton et al., 2016). Ecosystem-based measures also have an implicit benefit in that they generally do not harm the coastal environment like other 'hard' engineering structures (Bulleri, 2010; Gittman, et al., 2016). Estimating the economic value of these co-benefits is crucial when comparing ecosystem-based measures with other coastal adaptation measures.

4.4.4.2.5 Drawbacks

Similar to any other feature that interacts with coastal processes natural wetlands and reefs can increase flooding in some instances. This can happen, for example, due to the redistribution or acceleration of flows in channels within a wetland system (Marsooli, et al., 2016) or an increase in infragravity wave energy behind a reef (Roeber and Bricker, 2015). Understanding these effects is as important as evaluating their benefits when implementing ecosystem-based measures.

4.4.4.3 Advance, Accommodation and Retreat Measures

Contrary to protection measures, little systematic monetary information is available about costs, benefits and drawbacks of advance, accommodation and retreat measures, specifically not in the peer-reviewed literature. The costs of land reclamation are extremely variable and will depend on the unit cost of fill versus the volumetric requirement to raise the land. Hence, filling shallow areas is preferred on a cost basis. Benefit to cost ratios of land reclamation can be very high in urban areas due to high land and real-estate prices (Hinkel, 2018). The major drawback include groundwater salinisation, enhanced erosion and loss of coastal ecosystem (Li, et al., 2014; Nadzir, et al., 2014; W. Wang, et al., 2014; Chee, 2017). With regards to physical accommodation, cost-benefit estimates have been collected in the USA based on the National Flood Insurance Program, although this only addresses present extremes and ignores sea level rise (Linham and Nicholls, 2010). In this context, it has been estimated that elevating new houses by 2 feet might raise mortgage payments by USD240 a year, but reduce flood insurance by USD1,000 to USD2000 a year depending on the flood zone (FEMA, 2018). For Ho Chi Minh City, elevating areas at high risk and retrofitting buildings would have 21st century benefit-cost ratios of 15 under SLR of 180 cm and a discount rate of 5% (Scussolini, 2017). The economics of retreat differs from the ones of protect, advance and accommodation measures. Once implemented, maintenance costs hardly accrue. Assessments of few data available have estimated that monetary cost of managed retreat may vary between USD10,000 and 100,000 per person (Hino, et al., 2017).

4.4.4.5 Trade-offs, Synergies and Economic Efficiency Across Measures

So far little economic information is available on trade-offs, synergies and the overall efficiency across measures. A couple of new global studies confirm the findings of AR5 that protection against increased coastal flooding is economically efficient for urban areas during the 21st century (Wong, 2014). Amongst these studies there is *high agreement* that during the 21st century the benefits of reducing coastal flood risk through hard protection exceed the capital and maintenance costs for protection infrastructure for cities and densely populated areas, even under high-end sea level rise (Diaz, 2016; Lincke, 2018). For example, Diaz (2016) find that under 21st century SLR of 0.3 to 1.3 m and SSP2, adaptation in terms of protection and retreat reduces global net present costs of SLR by a factor of seven as compared to no adaptation, when applying a discount rate of 4%. ind that during the 21st century it is economically efficient to protect 13% of the global coastline, which corresponds to 90% of global floodplain population, under SLR scenarios from 0.3 to 2.0 m, five Shared Socio-economic Pathways (SSPs) and discount rates up to 6% (Figure 4.13).

These findings suggest that it generally makes economic sense to continue to protect existing urban areas by hard defences, following current practice (Section 4.4.2). High-end sea level rise above 2 m and beyond 2100 could certainly change this picture, but given the deep uncertainty involved, the point in time when it would be economically meaningful for cities to switch from a protection to a retreat strategy is still far away (Hinkel, 2018), although studies addressing specifically this question are currently not available. For less densely populated and currently not protected areas, the situation is more complex. For example, across the wide range of scenarios considered, Lincke (2018) find that for about 65% of the world's coast it is economically efficient not to protect and for 21% of the coasts no robust solution emerges (Figure 4.13). There is, however, no study available that has looked at robustness and economic efficiencies across all four types of response measures.

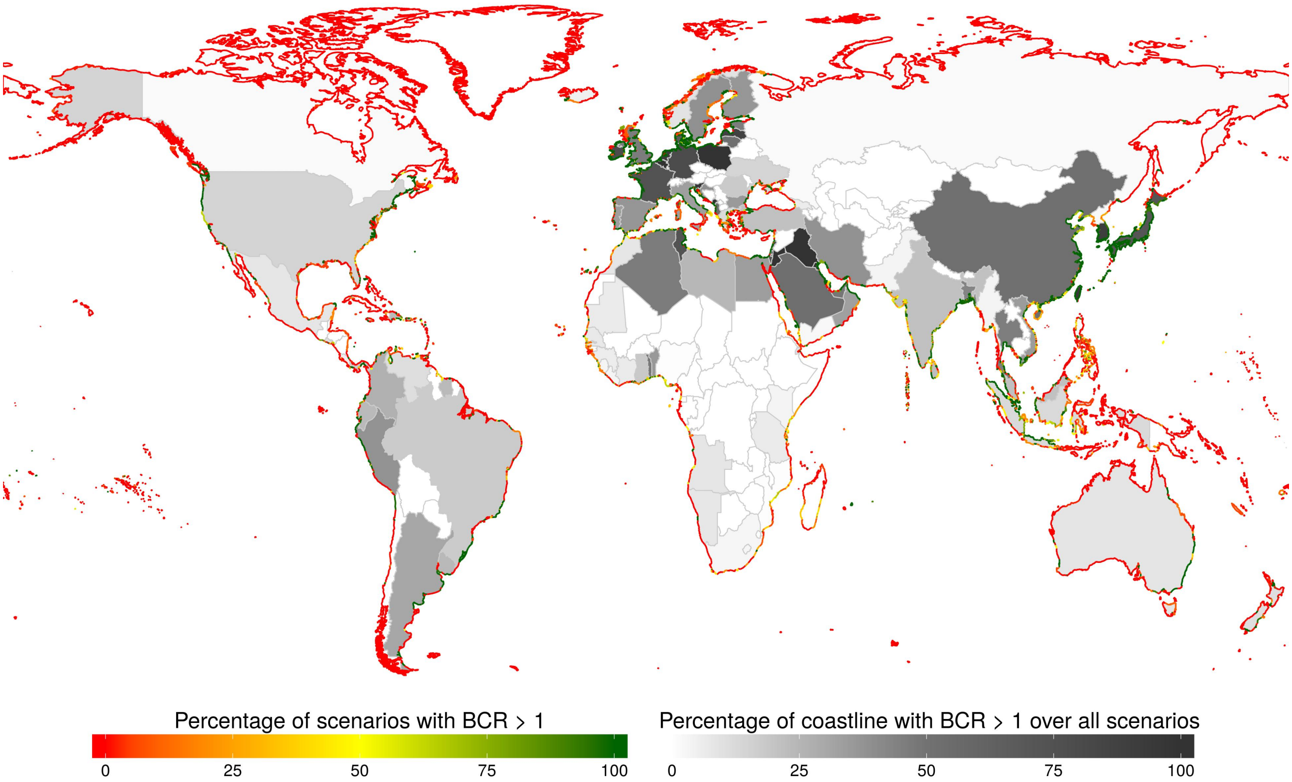


Figure 4.13: Economic robustness of coastal protection under SLR scenarios from 0.3 m to 2.0 m, the five SSPs and discount rates of up to 6%. Coastlines are coloured according to the percentage of scenarios under which benefit-cost ratio are above 1 and countries are shaded grey according to the share of their coastline having a benefit-cost ratio above 1. Source: (Lincke, 2018).

When designing responses, many more consideration can be made beyond those that have been considered in the quantitative studies reported above. Generally, there is *high confidence* regarding design considerations for hard protection and sediment-based measures. Beaches and dunes are well-established within coastal engineering and adaptation practice and there is *high evidence* and *high agreement* for their design process and considerations (Thieler, et al., 2000; USACE, 2002; De Jong, 2014). Hard protection measures are generally favoured where large human values are at risk such as in and around major coastal settlements. In principle, both beach-dune systems and dikes can be designed to deliver similar levels of flood risk reduction at the time of construction. With sea level rise (and potentially changes in storminess), the residual risk may increase differentially, depending on the approach to adaptation. With hard defences, adaptation involves making a series of increases in their crest elevation and associated cross section. With beach-dune systems, a more continuous monitoring and nourishment process is possible. In situations where the coastline exhibits significant changes in direction or where wave attack is strongly oblique, nourishment may need to be associated with hard structures to control alongshore transport of sediment (e.g., groins) or to reduce wave action to limit the required height and volume of beach material (e.g., offshore cills or breakwaters).

There is *medium evidence and low agreement* regarding design considerations for ecosystem-based measures. Critical gaps remain in our understanding of a range of factors that together affect the success of ecosystem-based measures including choice of species and restoration techniques, lead time, natural variability and residual risk, temperature, salinity, wave energy and tidal range (Smith, 2006; Stiles Jr, 2006). For example, common reasons for the failure of mangrove restoration projects include poor choice of mangrove species, planting in the wrong tidal zones and, planting in areas of excessive wave energy

(Primavera and Esteban, 2008; Bayraktarov, 2016; Kodikara, 2017). Furthermore, ecosystem-based measures may have differential lead times before they start providing coastal protection benefits which may necessitate intermediate defense measures. For example, newly planted mangroves will provide less wave attenuation until they mature (~3–5 years; Mazda et al., 1997). In contrast, a reef restoration project that uses submerged concrete structures will perform as a breakwater as soon as the sub-structure is in place (Reguero, et al., 2018). Another insufficiently understood design consideration for EBM is seasonal, annual and longer-term variability. For example, marsh and seagrass wetlands typically have lower densities in winter which reduces their coastal protection capacity (Paul and Amos, 2011). In the long-term, there is *limited evidence* and *low agreement* for how changes in sea level, sediment inputs, ocean temperature and ocean acidity will influence the extent, distribution and health of marsh and mangrove wetlands, coral reefs and oyster reefs (Hoegh-Guldberg, 2007; Lovelock, et al., 2015; Crosby, 2016; Albert, 2017).

In any case, EBM require more space to achieve the same level of protection than hard protection measure, and both high land costs and permitting costs can be barriers to EBS implementation in many geographies – and in some cases, with biases towards hard structures. Hence, EBM play a smaller role in densely populated urban areas. Combining EBM and hard protection is a promising way forward (Spalding, et al., 2014). Careful design of hybrid approaches can provide enhanced coastal protection while still providing a number of ecosystem services such as improved water quality (Sutton-Grier et al., 2015; Liquete et al., 2016). Most of these hybrid solution are very recent thus there are few data on their effectiveness or on the cost to benefit ratio (Pontee et al., 2016).

Unlike hard protection measures, which are fixed in position and height and will degrade in their risk reduction capacity without maintenance, ecosystem-based measures can improve their risk reduction capacity over time (van Wesenbeeck, et al., 2016), can naturally adapt to rising sea levels (Rodriguez, et al., 2014; Kirwan, et al., 2016; Woodroffe, et al., 2016), and recover from extreme events (Long et al., 2016). For example, oyster reefs have been shown to be able to keep pace with predicted sea level rise through 2100 (Rodriguez, et al., 2014).

Finally, it is important to note that protection, advance and accommodation measures are always associated with a residual risk, that is there is a finite probability of failure, however small. Only retreat measure can avoid these residual risks.

In line with AR5, it can be reiterated that the relevance of all of these measures, whether they are protection, advance, accommodate or retreat measures, is critically context-specific, depending on both natural and human circumstances.

[PLACEHOLDER FOR SECOND ORDER DRAFT: migration modelling for sea level rise driver-exposure, gravity, and history-based modelling]

4.4.5 Approaches for Making Social Choices and Appraising and Institutionalizing Adaptation Pathways

A diversity of methods and tools are available and applied for developing and appraising adaptation pathways. Approaches include decision-making methods, methods for facilitating social choice and transformation responses. The literature on this has rapidly developed since AR5.

4.4.5.1 Key Principles for Developing, Appraising and Institutionalizing Adaptation Pathways

[PLACEHOLDER FOR SECOND ORDER DRAFT: foundational principles for enabling adaptation as of AR5 to be summarised and post-AR5 literature regarding principles for adaptation to be assessed, with a focus on application of the adaptation pathways approach]

4.4.5.2 Decision Analysis

[PLACEHOLDER FOR SECOND ORDER DRAFT: application of adaptation principles to be introduced, starting with an succinct statement about decision analysis and assessment of SLR response, with focus on assessing post-AR5 literature]

4.4.5.3 Formal Approaches and Methods for Assessing Adaptation Options

4.4.5.3.1 Introduction

A range of formal, decision-analytical methods are available and applied for appraising and choosing adaptation options. AR5 gave an overview of available methods for adaptation generally (Chambwera, 2014; Jones, 2014), as well as for coastal adaptation specifically (Wong, 2014). Since AR5 the literature on formal coastal decision analysis has grown significantly. This section assesses recent advances.

Decision-analytical methods identify options (also called alternatives or adaptation pathways) from a predefined set of available options that perform best or well with regards to given objectives. An option is a specific combination of adaptation measures applied over time (See Section 4.4.2). Each option is characterised for each possible future state-of-the world (e.g., emission and socio-economic scenarios) by one or several attributes, which may measure any relevant social, ecological, or economic value associated with choosing and implementing the option (Kleindorfer, 1993). Attributes commonly used include cost of adaptation options, monetary and non-monetary benefits of the SLR impacts avoided, or net present value (NPV), which is the difference between discounted monetary benefits over time and discounted costs over time.

The formal analysis of decision does not suggest that there are purely objective ways of making decisions. The application of formal methods entails a number of normative choices concerning the objectives, the specific methods applied, the set of options considered and the attributes used to characterize options. Furthermore, adaptation decisions and their contexts are diverse and different context require different decision making approaches, with formal decision analysis being one approach next to other ways to inform social choices (Kleindorfer, 1993; Hinkel, 2016). Formal analysis is generally indicated if decisions are complex and target long term investments, as it is frequently the case in coastal context. But in any case, formal analysis of decisions can be embedded in a social process that ensures that the normative choices made reflect social needs and objectives. See also Sections 4.4.5.1 and 4.4.5.2.

Generally, adaptation options can be analysed against all available knowledge including all major uncertainties and also ambiguities amongst expert opinions and their distinct approaches, because considering uncertainty and ambiguity only partially may misguide the choice of adaptation options (Renn, 2008; Jones, 2014; Hinkel, 2016). In the case of coastal adaptation, it is thus necessary to consider uncertainty in global and regional mean and extreme sea levels (Sections 4.2.3.2, 4.2.3.5 and 4.2.3.4), waves (Section 4.2.3.4.2), local vertical land movement due to glacial-isostatic adjustment, tectonics and land subsidence (Section 4.2.2.5), as well as uncertainties in coastal impact modelling and socio-economic development (Section 4.3).

4.4.5.3.2 Maximisation of expected utility

The objective of cost-benefit analysis (CBA) and similar methods building on normative expected utility theory is to choose the adaptation option that has the lowest expected NPV (i.e., the cheapest option) or the highest expected social welfare (i.e., the socially efficient option). The application of CBA to compute an economically optimal height of a coastal protection structure has a long tradition dating back to the aftermath of the devastating coastal floods of 1953 in the Netherlands (van Dantzig, 1956). Without SLR, the application is straight forward. Each option is characterized by its costs and, under each state-of the world (here, extreme sea level), by its monetary benefits. Non-market benefits are taken into account through economic valuation (Section 1.1.4). Next, a probability distribution over the states-of-the world (e.g., distribution of extreme sea levels) is used to compute the expected NPV of each option. Finally, the option with the highest NPV is chosen. CBA comes along with several well-known limitations such its sensitivity to discount rates and the difficulty to monetize ecological, cultural and other intangible benefits that have been widely discussed in previous IPCC reports (Chambwera, 2014; Kunreuther, 2014). Specifically for the coastal context, uncertainties in model structure and parameters can result in large variations in the economically efficient solution (Oddo, 2017).

Climate change introduces the complication that information on future mean and extreme sea levels (as well as on any other climate variable) can only be attained contingent on a given GHG emission scenario. For example, under RCP8.5, more rapid sea level rise will lead to a larger increase in extreme sea levels than under RCP2.6. For the short term, emission scenario uncertainty is small and can be neglected, and hence CBA can be applied. Until when this is the case depends on the location, because both the relative and extreme sea level varies from place to place as discussed in Section 4.2.3.4.1 and in particular in Figure 4.9. If the variability in the tide gauge record is large with respect to the relative sea level rise it takes longer before the difference between scenarios becomes apparent in the annual maxima sea level. This is illustrated in Figure 4.14 which shows the year in which probability distribution of extreme sea levels for RCP2.6 and RCP8.5 start overlapping by less than 90% (termed year of scenario divergence here). For more than half of the coastal sites with sufficient observational data, this is the case before 2030, but for 4% of locations this occurs later than 2045. Beyond the year of scenario divergence, CBA can only be applied if subjective probabilities are attributed to emission scenarios and used to derive a single (i.e., emission scenario independent) probability distribution of future mean and extreme sea levels.

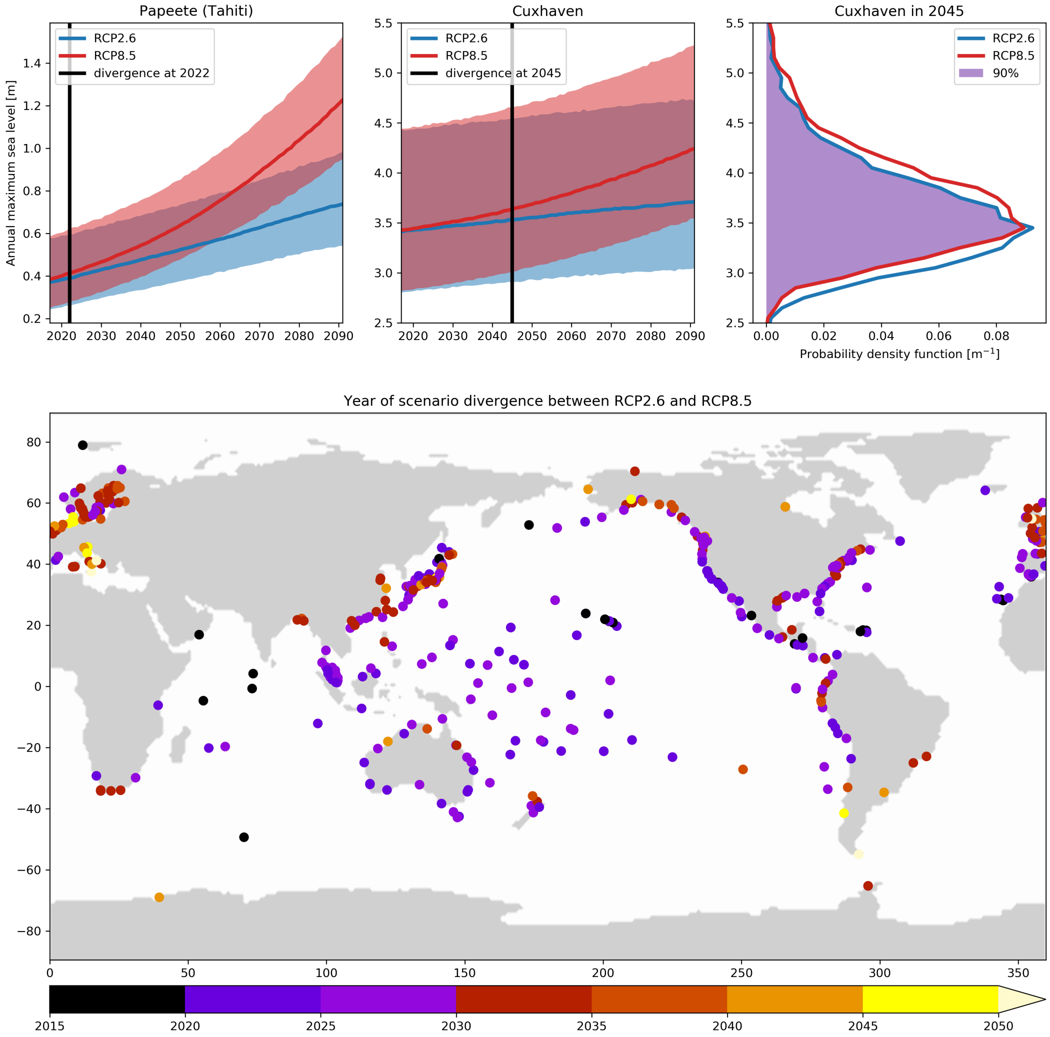


Figure 4.14: Scenario divergence for extreme sea level between RCP2.6 and RCP8.5. Upper left and middle panels indicate the median and 5–95 percentile range of future annual maximum sea level relative to the 1986–2005 baseline. Divergence is defined as the year where the probability functions for RCP2.6 and RCP8.5 overlap less than 90%, as illustrated in the upper right panel. The bottom panel indicates the year of scenario divergence for all coastal sites with sufficient observational data.

The question whether to attribute subjective probabilities to emission scenarios or not has been extensively discussed in previous IPCC reports (Carter, 2007). The main argument in favour of this is that decision makers are *likely* to misinterpret emission scenarios without probabilities as being equally *likely* (Schneider, 2001). Arguments against this include that space of possible future emissions is insufficiently sampled by any number of scenarios and individuals are *likely* to significantly disagree on subjective probabilities of emission scenarios (Lempert, 2001; Stirling, 2010). In the current SLR literature only very few cases can be found were probabilities have been attached to emission scenarios (Woodward, 2014; Abadie, 2018). Generally there is *high agreement* not to do so and hence not to apply utility optimisation methods such as CBA.

There is, however, extensive literature that applies scenario-based CBA, which consists in applying a separate CBA for each emission or SLR scenario considered. Arguably this is the case because CBA is generally widely used in public and infrastructure related decision making and also legally prescribed for coastal adaptation infrastructure projects in many countries such as the US, the UK and The Netherlands. For example, scenario-based CBA has been applied for setting the safety standards of Dutch dike rings (Kind, 2014; Eijgenraam, 2016), exploring future protection options for New York (Aerts, 2014), Ho Chi Minh City (Scussolini, 2017) and many other locations. While scenario-based CBA identifies an optimal option under each scenario, it does not formally address the problem a coastal decision maker is facing, namely to decide across scenarios (Lincke, 2018). Nevertheless, the results of scenario-based CBA provide some guidance for decisions and can also be used as inputs to robust and flexible decision making methods discussed in the next subsections.

4.4.5.3.3 Robust decision making

The objective of robust decision making (RDM) is to identify options that perform reasonably well (i.e., ‘robust’), under a wide range of future states-of-the-world. Used in a narrow sense, RDM refers to exploratory modelling methods that rely on simulation models to create large ensemble of plausible future scenarios for each option and then use search and visualization techniques to extract robust options (Lempert, 2000). A range of attributes such as costs, benefits, NPV (derived by scenario-based CBA as discussed above), regret, flexibility (see next Section), reversibility, security margins, etc. may be used to characterise options and define robustness criteria (Hallegatte, 2009). Used in a wider sense, RDM also includes methods that follow similar ideas such as minimax or minimax regret (Savage, 1951) or info gap theory (Ben-Haim, 2006).

RDM methods are particular suitable for coastal adaptation decision making for two reasons. First, they address the problem of deep uncertainty, that is the situation that no unambiguous probability distribution can be attached to states-of-the-world (Lempert, 2001; Weaver, 2013), which is the case for longer term SLR decision making as discussed above. Second, even if a probability distribution is available, these methods may be preferred over optimisation methods if decision makers are risk or uncertainty averse, which is frequently the case for the coastal context (Hinkel, 2015). Despite their suitability for coastal adaptation decisions, few applications are in the literature. For example Brekelmans (2012) minimize the average and maximum regret across a range of SLR scenarios for dike rings and Lempert (2013) apply RDM in Hoh-Chi-Minh City**.**

4.4.5.3.4 Flexible decision making

The objective of flexible decision making is to keep future options open by favouring flexible options over non-flexible ones. An option is said to be ‘flexible’ if it allows switching to other options once the implemented option is no longer effective. For example, a flexible protection approach would be to build small dikes on foundations designed for higher dikes, in order to be able to raise dikes in the future when more is known about SLR. Such staged approach is generally suitable for coastal adaptation due to the long lead and life-times of many coastal adaptation measures and the deep uncertainties in future sea levels (Hallegatte, 2009; Kelly, 2015).

A prominent and lightweight method that addresses the objective of flexibility is adaptation pathways analysis illustrated in Figure 4.15 (Haasnoot, 2011; Haasnoot, 2012). The method graphically represents alternative combinations of measures over time together with information until when options are effective (called adaption tipping point) under all (drawn through lines) and some scenarios (dashed-lines), as well as possible alternative options then available. A completed adaptation pathways plan thus suggest policy actions for the short to medium term, within a longer-term pathway. An adaption pathway starts with the current policy (called current situation in Figure 4.15 or basic policy by Walker, et al. (2013)). As time and SLR progresses, monitoring may trigger a new decision (depicted as a decision node) to select and prepare for switching to an alternative option.

Adaptation pathway analysis has been widely applied both in the scientific literature as well as in practical cases. Prominent applications after AR5 include Indonesia (Butler, et al., 2014), New York City (Rosenzweig and Solecki, 2014) and Singapore (Buurman and Babovic, 2016). The method has proven to be specifically useful in interaction with decision-makers and other stakeholders, helping them to identify appropriate sequences of measures over time, avoiding lock-ins and identifying path dependencies (Haasnoot, 2012; Haasnoot, 2013). Furthermore, the method has been valued highly because it illustrates decision-makers that there are several possible pathways of reaching a desired future (Haasnoot, 2013; Brown, et al., 2014; Werners, et al., 2015).

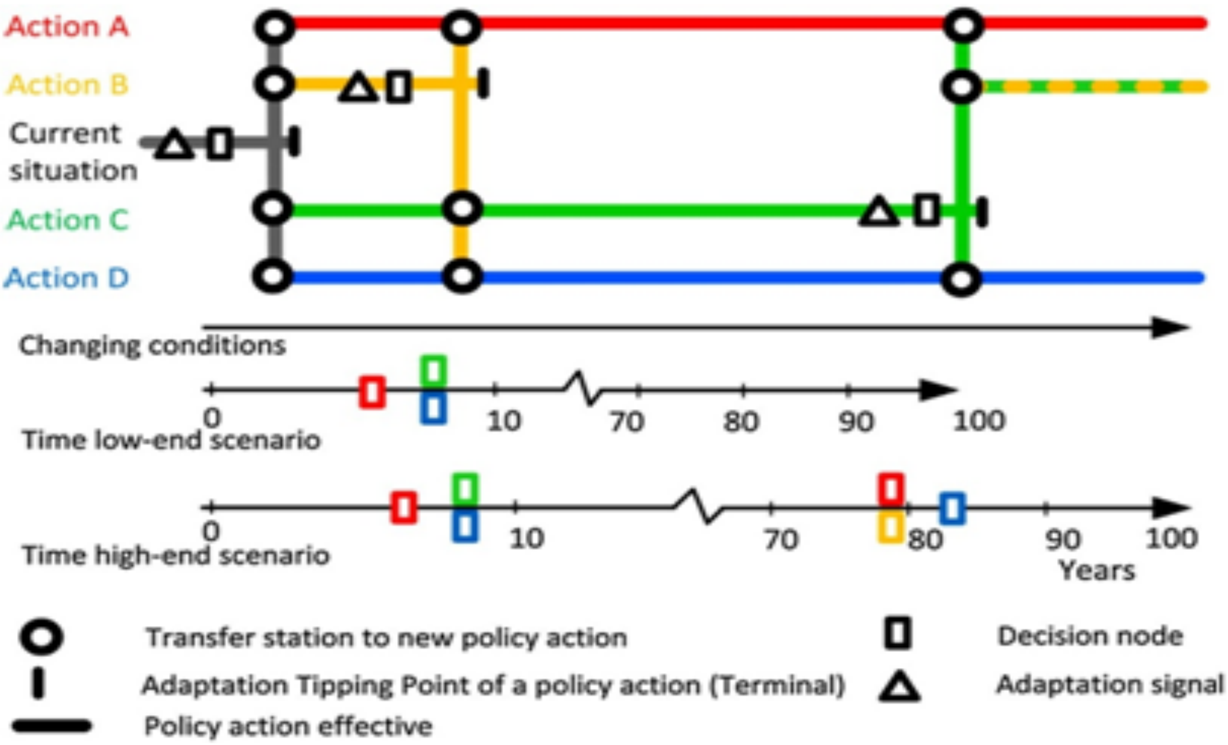


Figure 4.15: Key concepts in adaptation pathways approaches (Haasnoot et al., 2013).

While adaptation pathways analysis identifies adaptation pathway in terms of their flexibility, the method cannot answer the question of economically efficient flexibility and timing of adaptation. Delaying decisions and opting for flexible measures may introduce extra costs, because flexible measures are often more expensive than inflexible ones and flood damages may occur whilst delaying the decision. An important consideration therefore is to balance the cost of delaying decisions with the benefits of deciding later when having more information at hand. This is precisely the decision problem methods such as real-options analysis [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added: Dixit and Pindyck, 1994] and decision tree analysis [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added: Conrad, 1980] can address. The former is an extension of CBA and thus can only be applied when probabilities for emission scenarios are available as discussed in Section 4.4.5.3.2. A second requirement for both methods is that the reduction of future uncertainty can be quantified. Application of these approaches are few in the SLR literature. Woodward (2014) apply real-option analysis to an area of the Thames Estuary in London, England, and Buurman and Babovic (2016) for the cases of Singapore.

4.4.5.3.5 Further methods and research needs

The three broad categories of formal decision-making approaches organised above in terms of the three objectives of optimal expected utility, robustness and flexibility cover the main single-attribute decision-analytical methods applicable to coastal adaptation. Another category of approaches called multi-criteria analysis (comprising multiple-objective and multiple-attribute decision making) is also suitable for coastal adaptation, because adaptation often involves stakeholders having distinct objectives and valuing options differently (Oddo, 2017). Furthermore, methods are combined across categories. For example, in adaption pathways analysis options can also be characterized through multiple attributes such as costs, effectiveness, co-benefits, etc., with in turn can be used in a multi-attribute decision making method (Haasnoot, 2013). Other coastal examples include a combination of adaptation pathway and adaptive policymaking applied to the lower Rhine Delta in the Netherlands (Haasnoot, 2013) and a combination of adaptation pathways, adaptive planning and real-option analysis applied to coastal adaptation in Singapore (Buurman and Babovic, 2016).

Furthermore, there is wide agreement that next to making robust and flexible decision today, any decision making method can be applied within an adaptive and iterative policy cycle that includes monitoring and evaluation of options and sea level variables in order to learn from past decisions and collect information to inform future decisions (Haasnoot, 2013; Barnett, et al., 2014; Burch, et al., 2014; Wise, et al., 2014; Kelly, 2015; Lawrence and Haasnoot, 2017). Importantly, a monitoring strategy can be in place that helps to identify needed shifts in policy sufficiently upstream to limit the risk of negative impacts (Hermans, et al., 2017).

Three general gaps can be identified in the literature. First, the production of sea level rise information is insufficiently coupled to the use of this information in decision-analysis. This constitutes a limitation, as different coastal decision contexts require different decision-analytical methods, which in turn require different sea level rise information. Specifically, applications of decision-analytical methods generally convert existing sea level information to fit their method, often misinterpreting the information, making arbitrary assumptions or loosing essential information on the way (Hinkel et al., 2015; Bakker et al., 2017a; Van der Pol, 2018). Second, with the exception of adaptation pathway analysis, methods of robust and flexible decision making are under-represented in the literature despite their specific suitability (Van der Pol and Hinkel, 2018). Most of the literature focuses on scenario-based CBA, which does not address the problem of deep uncertainty coastal decision makers are facing. More research and applications of the technically advanced robust and flexible decision making methods that address multiple objectives is necessary. Third, research is necessary to better compare the various methods, to identity which methods are best suitable in which situation and to develop consist categorisations of methods. Available categorisations of decision-analytical methods for climate adaptation categorize methods differently (Hallegatte, 2012; Watkiss, 2015; Dittrich, 2016), so efforts to harmonize the field and strengthen cooperation between SLR physics and decision making would be highly beneficial (Hinkel, 2015).

An underlying challenge is to design and integrate relevant formal decision-making approaches into the heterogeneous reality of local planning and decision-making cultures, institutions, processes and practices, often with community-specific needs and requirements.

4.4.5.4 Community-based Approaches and Methods

This sub-section starts with a synopsis of AR5 understanding about community-based adaptation approaches and then assesses post-AR5 literature. Note that many of the approaches and methods discussed here have application in both developed and developing country contexts. Moreover, community engagement is invariably an important and integral part of processes that apply the formal assessment methodologies outlined above.

4.4.5.4.1 Reflection on relevant findings of AR5

[PLACEHOLDER FOR SECOND ORDER DRAFT: relevant citations added; albeit a summary of findings as at AR5 plus SR1.5]

AR5 noted that, as the complexity of management challenges increases due to climate change, development and other pressures, community-based adaptation (CB) and other bottom-up, reflexive decision-making processes have emerged over the last few decades or so. For example, coastal adaptation now involves a wide range of approaches and frameworks, including CB along with integrated coastal management, ecosystem-based adaptation, community-based adaptation, and disaster risk reduction and management.

CB is also one of the emerging, integrated approaches to adaptation planning, governance, and implementation, with these being viewed as more effective than standalone efforts to reduce climate-related risks. Such integration is important, as many sectors experience threats not only from climate change, but also from a range of existing or emerging threats (see previous Sections).

In parallel with national-level planning, CB has become an increasingly prevalent practice, particularly in developing counties. Where a combination of top-down and bottom-up activities has been undertaken, the links between adaptation planning and implementation have been strengthened. In either approach, participation by a broad spectrum of stakeholders, and close collaboration between research and management, have been identified as important mechanisms to undertake and inform adaptation planning and implementation.

It is increasingly apparent that CB potentially offers ways to address the vulnerability of local communities by connecting climate change adaptation to non-climate local needs. CB approaches have been developed through active participatory processes with local stakeholders, and operated on a learning-by-doing, bottom-up, empowerment paradigm. For example, with the social dimensions of climate change adaptation receiving more attention, there has been an increased emphasis on addressing the needs of the groups most vulnerable to climate change, such as children, the elderly, disabled, and poor. Increased capacity, voice, and influence of low-income groups and vulnerable communities and their partnerships with local governments, have been shown to benefit adaptation.

Given the diverse physical and human attributes of small islands, CB has been shown to generate larger benefits when delivered in conjunction with other development activities. The relevance of CB principles to island communities, as a facilitating factor in adaptation planning and implementation, has been highlighted, for example, with focus on empowerment and learning-by-doing, while addressing local priorities and building on local knowledge and capacity. CB can include measures that cut across sectors and technological, social, and institutional processes, recognizing that technology by itself is only one component of successful adaptation. But capacity barriers have hampered the transition from planning to implementation. Participatory consultations across stakeholders and sectors within communities and capacity building taking into account traditional practices can be vital to the success of adaptation initiatives in island communities.

CB can support transformation where it engages with key development agendas to reduce poverty and vulnerability, and can address local inequalities and adverse power relations at district, city, national, and transnational levels. But urban governance regimes are often resistant to change, and civil society organizations can be marginalized or co-opted, thus reducing the scope for transformative adaptation.

CB involves local people in a participatory and collaborative manner through the merging of scientific and local knowledge, to improve resilience and ensure sustainability of adaptation plans. CB has been criticised for not always representing vulnerable people fairly and for failing to build long-term social resilience, it highlights that evidence from climate change-affected communities indicates that CB provides benefits by increasing local adaptive capacity in order to improve livelihood assets and security, as well as addressing inequalities and gender biases at the local level.

Although engineering and technological measures currently dominate adaptation efforts, ecosystem-based, community-based, and institutional and social approaches are increasing. To effectively adapt to climate change, bottom-up initiatives by individuals and communities are essential, in addition to efforts of governments, organisations, and institutions. Depending upon the context and vulnerability of specific communities, community-based adaptation can be an effective adaptation option.

The importance of community participation for mitigation and adaptation is highlighted, and in particular the need to take into account equity and gender considerations. But community participation also brings challenges, and may not always result in better policy outcomes. Stakeholders, for example, may not view climate change as a priority and may not share the same preferences, potentially creating policy deadlock.

Fundamentally voluntary actions by non-governmental actors are gaining importance and could make an important contribution. Some studies suggest that better long-term results are achieved through interpersonal or community-based initiatives. While adaptation initiatives by individuals may temporarily reduce the impacts of climate change and allow residents to cope with changing environmental circumstances, they may not be sufficient to sustain communities' way of life in the long term. Therefore, more long-term and sustainable adaptation initiatives are needed. Thus, to achieve successful long-term adaptation, integration of individuals' adaptation initiatives with top-down adaptation policy will be critical. Failing to do so may lead individual actors to mistrust authority and can discourage them from undertaking adequate adaptive actions.

However, it is challenging to mainstream CB into national and local planning. CB that is grounded on community values, coping strategies and decision-making structures cannot work in isolation at the community level since factors beyond the control of the community scale, such as governance and policy context, affect their vulnerability to climate change.

With CB it is often difficult to uphold principles of equity, justice and ensure access to information that is fair for all. Dominant or normative pathways tend to validate the practices, visions, and values of existing governance regimes and the more privileged members of a community, given their assets and long-standing power positions, while devaluing those of less well-off households, different ethnic groups, and other disenfranchised stakeholders, thereby exacerbating inequalities and pushing the most vulnerable toward lock-in situations with less and less capacity to navigate change.

Tensions between values and worldviews that influence CB pathway decisions, such as individual economic gains and prosperity versus community cohesion and solidarity, further erode collective adaptive action. Moreover, innovative actions that deviate from the dominant path are discouraged. A narrow view of decision making, for example focused on technical feasibility and cost-benefit analyses, tends to crowd out more participatory and inclusive processes that underpin collective learning and wider consultation, and can obscure contested values and power asymmetries in governance.

A situated and context-specific understanding of place that brings to the fore multiple knowledge sources, values, and contested politics helps to overcome dominant path dependencies, challenge scientific options detached from place, and advance joint place making. Such understanding suggests that win-win outcomes, even via socially-inclusive adaptation pathway approaches, will be exceedingly difficult to achieve without a commitment to inclusiveness, place-specific trade-off deliberations, redistributive measures, and procedural justice mechanisms to facilitate equitable transformation, including achieving poverty eradiation and reducing inequalities.

Community-led and bottom-up approaches offer potentials for climate-resilient development pathways at scale. At the level of individuals, communities, and groups, emphasis on well-being, social inclusion, equity, and human rights helps to overcome limitations in capacity. Social influence approaches that do not involve social interaction, such as social norm, social comparison and group feedback, are less effective, but can be easily administered on a large scale at low cost.

While participatory governance and iterative social learning constitute key aspects to enable transformative social change and climate resilient development pathways, dominant pathways and entrenched power differentials continue to undermine the rights, values, and priorities of disadvantaged populations in decision making. Community-level climate resilient development pathways that focus on capabilities and capacities can provide an important complement to national trajectories, flagging potential negative impacts of state-level commitments on disadvantaged groups, such as low-income communities and communities of colour. They underscore the crucial roles of social equity, participatory governance, social inclusion, and human rights, as well as innovation, experimentation, and collective learning.

This approach to Common but Differentiated Responsibilities and Respective Capacities', as defined in the UNFCCC, implies choosing climate actions that create opportunities and benefits and allow people to live a life in dignity while avoiding actions that undermine capabilities and erode well-being. It is in alignment with transformative social development, and the 2030 Agenda of "leaving no one behind", aiming to preclude severe limitations in adaptive capacities while supporting transformation and strengthening resilience.

4.4.5.4.2 Moving beyond AR5 in the context of CBA to SLR in low-lying islands, coasts and communities

Participatory approaches and tools

Communities have many opportunities to use participatory approaches when deciding how best to respond to climate change. Such approaches are most effective when designed and planned as a long-term-process that: (1) empowers people to handle by themselves the challenges and influence the direction of their interventions, leading to joint decision-making about what can be achieved and how; and (2) uses the inputs and opinions of stakeholders in the community to achieve an externally defined pre-established goal as by the national government.

Participatory approaches have been used by communities in the following areas: needs assessment, design and management of protective barriers, monitoring sea level changes, teaching and training on the basics of climate change and its impacts on the coasts, research and conflict management. These areas involved a range of stakeholders (e.g., multi-level workshops, for example: public sector senior managers, practitioners and community members, youth, parents, senior citizens, disabled citizens, diverse ethnic backgrounds, and higher education staff and students) involving sharing of knowledge and experience in coastal projects, working in teams on practical tasks, the use of visualization and analytical tools, and the development of shared understanding of climate change specifically, sea level rise and their implications to coastal development. Hence, participatory approaches take the form of simply being informed (passive participation), or answers are provided stakeholders (participation by consultation), or participation in the discussion and analysis of predetermined objectives set by the community (participation by collaboration), or primary stakeholders initiate the process and take part in the analysis and evaluation, leading to joint decision-making about (empowerment participation; Fortes, 2018).

In the context of community participation in Bangladesh to cope with sea level rise, and to enhance community resilient capacity, HAQUE et al. (2016) show that local-level CB approach and community involvement in decision-making processes are very effective in resilience building among the most vulnerable segment of the society. Community participation can be integrated in the broader national strategies for developing effective adaptation as well as social- ecological resilience system, while multilevel social networks are essential for developing social capital for supporting the legal, political, and financial frameworks that enhance community resilience. Integration of indigenous knowledge and learning from local communities into wider national policies can help ensure pro-poor climate governance.

While based on a review of community-based adaptation research in the Canadian Arctic, there are wider implications for the warning of Ford et al. (2016b) against assuming that research has a positive role to play in community adaptation just because it utilizes participatory approaches. They note that participation in CB research can perpetuate the privilege of Western knowledge over local values and indigenous knowledge, and can further marginalize communities if power relations are not addressed. Moreover, as CB also does not necessarily prevent maladaptation.

Hardy et al. (2017) argue that treating sea level rise as a social-ecological phenomenon, rather than just as a physical or ecological problem, has the potential to overcome the barriers to engagement with underrepresented communities, including through race-aware adaptation planning that encourages discussions at the onset of project formation to include issues of power and racial inequalities. Such a focus on livelihoods and everyday lives would necessitate a more complex policy process including investigations into the historical conditions that led to uneven racial development by partnering with organizations from underrepresented communities and groups and bringing local knowledge into the research design and planning phases. The practice of race-aware adaptation planning offers pathways to resist "passive indifference" and inequalities that perpetuate differentiated vulnerability to sea level rise.
A risk-based and participatory approach was used to assess climate vulnerability and improve governance in coastal Uruguay. It included a stakeholder-driven Vulnerability Reduction Assessment and multicriteria approaches to adaptation within a participatory bottom-up and top-down process. Effective coastal adaptation in Uruguay requires that technical knowledge be merged with lessons learnt through an adaptive management cycle to meet both short-term decision objectives and long-term adaptation goals (Nagy et al., 2015).

Broto et al. (2015) offer an appraisal of participatory urban planning for adaptation in practice, building upon a participatory experience in Maputo, a coastal city in Mozambique. They found that such bottom-up planning may lead to a more inclusive, and potentially fairer, society. However, the timescales of development are longer than those of participatory urban planning. This may lead to loss of momentum within the community. Moreover, participatory planning processes require a partnership built on mutual trust and understanding between local institutions and communities. An appropriate process of institutional support needs to be in place, but local governments often have difficulty integrating climate change knowledge as well as delivering adaptation interventions.

Garschagen (Submitted) used a participatory scenario-based approach to decipher the adaptation-development-nexus and its potential future trajectories for four coastal megacities. The approach was useful in convening a transdisciplinary learning and reflection process that shed light not only on the enablers but also barriers of transformation. Participatory Three-Dimensional Modelling (P3DM) has helped integrate indigenous and scientific knowledge systems (Piccolella, 2013). It added credibility to locally produced content and provided a platform for multi-stakeholder dialogue, while minimising the risk that perverse power dynamics would jeopardise the effectiveness of the participatory process. The combination of the community discussion with demonstrations of 'what if' scenarios of sea level rise confirmed that P3DM is able to bridge top- down and bottom-up approaches to coastal CBA by creating a space for mutual learning. Treuer (2017a) reports the findings of an immersive simulation experiment that accelerated 348 South Florida homeowners through thirty-five years of sea level rise. Levels of concern, and willingness to move, increased with higher sea levels, and was consistent across age, income, political identity, and other demographic divisions.

Community visioning and pathways

To overcome difficulties to build public support for flood-related policy and action, adaptation scenarios and 3D landscape visualizations were used in a visioning process to explore a range of alternative response options to sea level rise for the 100,000 citizens of Delta, a low-lying municipality at the mouth of the Fraser River delta. A large portion of the community is at considerable risk from climate change induced sea level rise and storm surges. The findings were used to support decision-making and further policy development for flood management in the municipality (Barron, undated).

Kench et al. (2018) challenge existing narratives of island loss by showing that island expansion has been the most common physical alteration throughout Tuvalu over the past four decades, despite sea level rising. The results are used to project future landform availability and consider opportunities for a vastly more nuanced and creative set of adaptation pathways for atoll nations.

Residents of a small town in coastal Australia used their photographs and accompanying narratives to both vision (by elucidating their current experiences) and re-envision (in advocating for different futures) their everyday experiences of adapting to flooding. The photoelicitation process provided different outcomes to conventional interviews, focus groups and questionnaires (O'Neill and Graham, 2016).

Brown et al. (2017) explored the lessons of coastal planning and development for the implementation of proactive adaptation in order to identify and open up windows of opportunity in current decision-making, to better design and implement proactive adaptation. They found that the windows of opportunity concept can aid practitioners and policymakers to identify instances where decision-making can be reframed or transformed to better enable proactive adaptation. Reframing of existing policies or creation of new transformative approaches can help build both the social capital and practical mechanisms required to deliver proactive adaptation. The transition from current to proactive adaptation requires shifts in decision framing, the pre-conditions and processes and outcomes association with identifying and opening adaptation spaces. Identifying windows of opportunity and understanding how they operate can support sustainability and adaptation mainstreaming and dynamic adaptation pathway approaches to help deliver the transformational change necessary for sustainable coastal adaptation to climate change.

Hay et al. (2018) call for increased understanding of the longer term habitability of atolls and islands. Changes in habitability occur as a result of the interplay between atmospheric, oceanic, social and economic conditions over the long-term. While a focus on resilience tends to favour responses that consider only the short-term, a longer term perspective is critical when considering strategic responses, such as international migration as an adaption option for countries facing severe declines in habitability. The drivers of declining habitability include increasing population density, economic vulnerability, and incidence of pests and disease.

Consensus building and decision making

Fortes (2018) advocates for constant dialogue among community and other relevant stakeholders in order to clarify the problem(s) to be addressed and subsequently initiating the iterative adaptive management cycle where problems are resolved through more informed decision making. Gorddard et al. (2016) argue that decision makers tend to frame adaptation as a decision problem, whereby the responses to impacts of change are addressed within existing decision processes centred on defining the decision problem and selecting options. As this approach is constrained by societal values and principles, regulations and norms and the state of knowledge, it is unsuitable for addressing complex, contested, cross-scale problems. But simply broadening the decision-making perspective to account for institutions and values is insufficient [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be added: Goddard et al. 2016]. When they analysed the influence of values, rules and knowledge on decision making and decision contexts for three adaptation projects that responded to sea level rise they found that linking these systems facilitates adaptation practitioners structuring adaptation as a process of co-evolutionary change that enables a broader set of social issues and change processes to be considered.

In a study of local adaptation planning processes in Vanuatu, Granderson (2017) found a marked difference in how actors contextualized and prioritized risks. Villagers assessed current impacts and risks from climate change in relation to wider socio-economic changes, and prioritized maintaining their way of life. In contrast, practitioners in civil society organizations (CSO) adopted a technocratic approach, drawing on climate science and focusing not only on the severity of risks but also on the potential need for external interventions. Explanations for climate-related changes, and notions of causality, also differed among villagers and CSO actors. Such differences in actors' values and politics highlight the need for open and inclusive dialogue that provides space for alternative understandings of risk, and for adaptation strategies that ensure community buy-in.

An Urgency, Barriers, and Risk (UBR) Framework assisted stakeholders in Miami Beach, Florida, to recognize and respond to barriers encountered in adapting to sea level rise (Treuer, 2017b). The Framework provides a structure for dynamically tracking and analysing the interaction between pressures driving policy change, decision makers, and barriers within the adaptation process. Three lessons were learned in its use: (1) Barriers to achieving consensus appeared towards the end of the agenda setting phase and early in implementation, when newly engaged stakeholders recognized and opposed specific adaptation actions; (2) Facilitation—based on the Netherlands approach to third-party facilitation in difficult, experimental climate adaptive water management projects—proved successful in overcoming barriers; and (3) Adaptation actions that address sea level rise risks at multiple timeframes were more successful.

Based on a study of a vulnerable, low-lying coastal area in northern Portugal, Campos et al. (2016) showed that participatory action research (PAR) was able to trigger new dynamics for collective decision-making that supported a sustainable direction in transformational adaptation. PAR uncovered the intricacies of planning and political processes, including the context-specific challenges for implementation. These include difficulties in translating decisions resulting from participative processes into effective policies. By building a support base from a wide group of stakeholders, PAR encourages socio-political legitimacy and trust for the results, such as the prioritization of adaptation options. PAR and transition management can complement each other in transition studies. By being more pragmatic, PAR can influence incrementally transformative changes that are guided by transition management's long-term design for governing sustainable transitions.

PRA projects in several coastal communities in New Brunswick and Quebec were found to deliver tangible short-term results as well as reinforcing the communities' governance and adaptation capacity and resilience over the long term. The engagement of stakeholders, and the exchange of information between scientists and local actors, led to a better evaluation of vulnerabilities and adaptation options. In some cases this resulted in the co-construction of new knowledge and the coproduction of priorities to build adaptation plans and tools with and for the communities. Thus, reflexive options such as sea walls were sometimes substituted by less costly and more targeted adaptation options, that are better suited to local circumstances and to the values and aspirations of the community. These solutions are more easily accepted within the community as well as by government authorities. However, not all projects led to immediate decision-making. The option of coastal retreat remains especially highly contentious and emotionally charged (Chouinard et al., 2015; Chouinard et al., 2017).

The foregoing confirm the SR1.5 conclusion that adaptation planning and interventions grounded on community values, coping strategies and decision-making structures cannot work in isolation at the community level since factors beyond the control of the community scale, such as governance and policy context, affect the ability to reduce vulnerability to climate change (*high evidence, high agreement*).

Learning from monitoring and evaluation
Mathew (2016) review and provide examples of community-based monitoring and evaluation initiatives, including participatory monitoring and evaluation. Such procedures help ensure increased authenticity of locally relevant findings and improve local capacity.

The preceding summary of AR5 and SR1.5 findings shows there are few examples of literature, understanding and application related to this topic. (Jhan, 2017) developed a modified Analysis-Awareness-Action framework to evaluate local climate change adaptation in four coastal townships along the vulnerable southwest coast of Taiwan in order to derive recommendations for local adaptation framework development. He found that a constructive dialogue and participatory processes are in order to increase community engagement in local adaptation. Improvements included engaging other local organisations and private actors, developing specialist organisations, legislative acts, and considering multiple objectives in formulation of adaptation actions to eliminate the potential conflict of interest.

The use of adaptation pathways implies a systematic monitoring effort to inform future adaptation decisions, with the monitoring feeding into a long-term collaborative learning process between multiple actors at various levels. Hermans et al. (2017) have developed an approach based around the conceptual core offered by adaptive policy pathways methods and their notion of signposts and triggers. This is embedded in a wider approach that revisits the critical assumptions in underlying basic policies, looks forward to future adaptation decisions, and incorporates reciprocity in the organization of monitoring and evaluation. The usefulness and practical feasibility of the approach were assessed using the Delta Programme in the Netherlands. This incorporated adaptation pathways in its adaptive delta management planning approach. The results suggest that the approach proposed by Hermans et al. (2017) adds value to existing monitoring practices. They also identified different types of signposts - technical signposts, in particular, need to be distinguished from political ones, and require different learning processes with different types of actors.

4.4.6 Barriers and Enablers and Lessons Learned for Adapting to SLR

[PLACEHOLDER FOR SECOND ORDER DRAFT: synopsis of relevant literature up to AR5 to be provided before delving into post-AR5 assessment that builds upon past literature on barriers to CCA in particular.]

4.4.6.1 Accounting for the Rate and Extent of Sea Level Rise

Considering rates of change affects the projected optimal adaptation strategy. Adaptation to a new climate state, instead of adaptation to ongoing rates of change, may produce inaccurate estimates of damages to the social systems and their ability to respond to external pressures. Shayegh et al. (2016) confirm this by determining the optimal investment taking into account the interplay among physical and economic factors governing coastal development decisions, including rate of sea level rise, land slope, discount rate, and depreciation rate. Optimal investment strategies depend on taking into account future rates of sea level rise, as well as social and political constraints.

Planners and decision makers who face risks arising from elevated sea levels are now provided with vastly improved hazard data and tools (see Section 4.2). While sea level rise is widely identified in adaptation plans for US coastal cities that are considered at high risk and vulnerable to rising sea levels, the overall quality of the plans to address it requires significant improvement (Fu et al., 2017). Localities lack the necessary information and incentives to plan for such an emerging agenda.

4.4.6.2 Accounting for Uncertainty

The process of decision making using adaptive pathways approaches is increasingly being used to plan for adaptation over time. It requires risk and uncertainty considerations to be transparent in the scenarios used in adaptive planning. In this regard, Stephens et al. (2017) have developed a framework for uncertainty identification and management within coastal hazard assessments. It can better inform identification of trigger points for adaptation pathways planning and their expected time range, compared to traditional coastal flooding hazard assessments.

In theory, application of adaptation pathways can also help ensure more implementable adaptation planning (Barnett et al., 2014).

However, developing pathways toward transformation is especially difficult in coastal regions where there are often multiple contested resource uses and rights, with diverse decision makers and rules, and where high uncertainty is generated by differences in stakeholders' values, understanding of future sea levels, and ways of adapting (Abel et al., 2016).

Those charged with developing and implementing coastal planning policies can recognise, communicate, and seek to overcome uncertainty. This can involve: 1) acknowledging and communicating uncertainties in existing and projected rates of sea level rise; 2) engaging in site-specific mapping based upon best available scientific information; 3) incorporating probabilities of extreme weather events; 4) resolving whether coastal engineering solutions are included in mapping; 5) ensuring that mapping includes areas required for future ecosystem migration; 6) managing discretion in planning and policy decision making processes; 7) creating flexible policies which can be updated in line with scientific developments; and 8) balancing the need for consistency with the ability to apply developments in science and technology (Bell et al., 2014).

A qualitative analysis of climate change adaptation initiatives in the small island nation of Kiribati revealed that adopting a culturally appropriate short-term (approximately 20 years) planning horizon may help reduce uncertainty and the trade-offs between adaptation options. In the short-term, the range of sea level projections may be small enough to not seriously confound adaptation decisions. But decisions can be regularly revisited, based on data collected on their effectiveness and reviews of the latest global sea level data and projections (Donner and Webber, 2014).

Thorarinsdottir et al. (2017) used two illustrative examples—Bergen on Norway's west coast and Esbjerg on the west coast of Denmark—to highlight how technical efforts to understand and quantify uncertainties in hydrologic projections can be coupled with concrete decision-problems framed by the needs of the end-users using statistical formulations. They found that failing to take uncertainty into account can result in the median-projected damage costs being an order of magnitude smaller.

The Adaptive Delta Management approach used in the Netherlands accommodates uncertainty in future climate and socio-economic changes. Key points of the approach are: 1) linking short-term decisions with long-term tasking around flood and drought risk management; 2) incorporating flexibility in possible adaptation strategies; 3) working with multiple adaptation strategies through adaptation pathways; and 4) linking different investment agendas. It provides greater transparency to decision-makers and stakeholders, as demonstrated for the management of flood risk and resilience in Dordrecht (Gersonius et al., 2016).

Construction of a consensus estimate from divergent expert assessments of rates of sea level rise can be subject to considerable structural (and deep) uncertainty (Bakker et al., 2017). As a result, a robust strategy, i.e., one that performs well over a wide range of plausible futures/views, may be preferable over optimal strategies. Effective communication of deep uncertainties depends strongly on the decision-context. Therefore, an efficient representation requires a tight interaction between decision analysts, scientists, and decision makers.

4.4.6.2.1 Barriers

The call for monitoring and evaluation of adaptation pathways stems first and foremost from the expectation that adaptation pathways are being implemented and that the developed plans help identify the variables that need to be monitored (Hermans et al., 2017). However, Wise et al. (2014) suggest that adaptation plans are often not implemented and, if they are, it is only the smaller incremental measures within those plans. van der Brugge and Roosjen (2015) explain how the presumed implementation and effectiveness of adaptation pathways might be due to the changes in institutional and socio-cultural structures required for the implementation of adaptation strategies. This implementation problem is not unique to adaptation pathways (Hermans et al., 2017).

4.4.6.2.2 Adaptive decision making

The AR5-cycle reported the emerging realization in the scientific and policy arena of the need to think adaptation and resilience policies and practices in a dynamic way (Brown et al., 2014), in order to 1) reflect the evolving nature of exposure and vulnerability (Denton et al., 2014), 2) improve the projection of climate change impacts (Cardona et al., 2012), 3) start anticipating the risks of maladaptation (Cardona et al., 2012; Noble et al., 2014), and (iv) enhance flexibility to allow better addressing climate change uncertainty (O'Brien et al., 2012; Noble et al., 2014). The 'adaptation pathway' approach thus gained attention, calling for 'cautious and staged implementation' (Kelly, 2015), for instance, long-term adaptation strategy based upon decision cycles that, over time, explore and sequence a set of possible actions based on alternative external, uncertain developments (Haasnoot, 2013; Barnett et al., 2014; Wise et al., 2014). The AR5-cycle recognizes the context-specific nature of adaptation pathways, that reflect 'competing prioritized values and objectives, and different visions of development that can change over time' (O'Brien et al., 2012, p. 440). Such a shift in adaptation thinking carries a real potential for a better integration of SLR and gradual changes more broadly. Until very recent works however, adaptation pathways have been described in a very general, theoretical way, with rare practical examples (Denton et al., 2014). The recent literature provides a better understanding of the dynamics of exposure and vulnerability, as well as first practical examples of adaptation pathways.

4.4.6.3 Accounting for Dynamics of Exposure and Vulnerability, and Path-dependency

4.4.6.3.1 Uncertainty manifestation in the policy system

In response to uncertain environmental and socio-economic change, those managing flood risk are urged to develop adaptive plans to ensure communities' long-term sustainable economic development (Hallegatte et al., 2016). However, there are challenges in developing and implementing such plans to address changing climate impacts and socio-economic conditions, including; dealing with uncertainty and the need to do so; understanding and acknowledging different types of uncertainty; making robust and adaptive decisions that can cope with uncertainties about the future, and shifting planning practice from static to dynamic approaches (Lawrence and Haasnoot, 2017).

4.4.6.3.2 Transformation, tipping points, and the acceleration of adaptation practice

Burch et al. (2017) discuss the emerging literature on transformative adaptation (Kates et al., 2012; Pelling et al., 2015) when paired with broader studies of development path shifts and sustainability transitions, suggests that the following are examples of actions or approaches that might push communities towards a tipping point' into a fundamentally more desirable state:

- Transformations in organizational methodology, major investments in capacity, new skills and ways of working (Pelling et al., 2015).
- Consideration of long-term (future) and irregular risks along with immediate risks (ibid).
- Engaging with a diversity of actors and interests, both within and among organizations (ibid) in a way that may trigger the imaginations of stakeholders and create excitement (Dempsey et al., 2011).
- Actions that do not simply consider key areas such as urban form, transportation, energy systems etc. but explicitly target the linkages among them (McCormick et al., 2013).
- Iterative, adaptive management that is based on monitoring and evaluation of key indicators (Burch et al., 2014).
- Framing climate change adaptation (or mitigation) in the context of sustainability/sustainable development, so as to capitalize on synergies and align the efforts with a wider variety of stakeholders (Shaw et al., 2014).
- These early suggestions of approaches that might lead to transformative adaptation reinforce the notion that transformation can be pursued through addressing the root causes of unsustainability, identifying tipping points that can act as leverage points, and by employing a social–ecological systems lens to reveal strategies that create synergies (and avoid trade-offs) with other priorities, Burch et al. (2017).

4.4.6.3.3 Potential for transformation

Also, Burch et al. (2017) were discussed the tipping toward transformation "progress, patterns and potential for climate change adaptation in the global south" In response to observed and projected climate change impacts, major donors are funding an abundance of climate change research. This is to capturing the broader trends and patterns across south cases. Furthermore, he recognizes that transformational approaches are difficult to implement for a variety of reasons. Kates et al. (2012) mention a key obstacle for transformative adaptation is uncertainty around the severity of climate impacts and purported benefits of adaptation. This includes uncertainty around the costs of transformation, which are often unknown but presumed to be high, compared to the ability to calculate the costs of incremental adaptation. Burch et al. (2017) also stated that: finally, there is a host of institutional barriers, ranging from cultural norms to existing complex legal systems. These barriers certainly exist in the countries in which the projects and adaptation options were conducted

4.4.6.4 Governance Barriers and Enablers

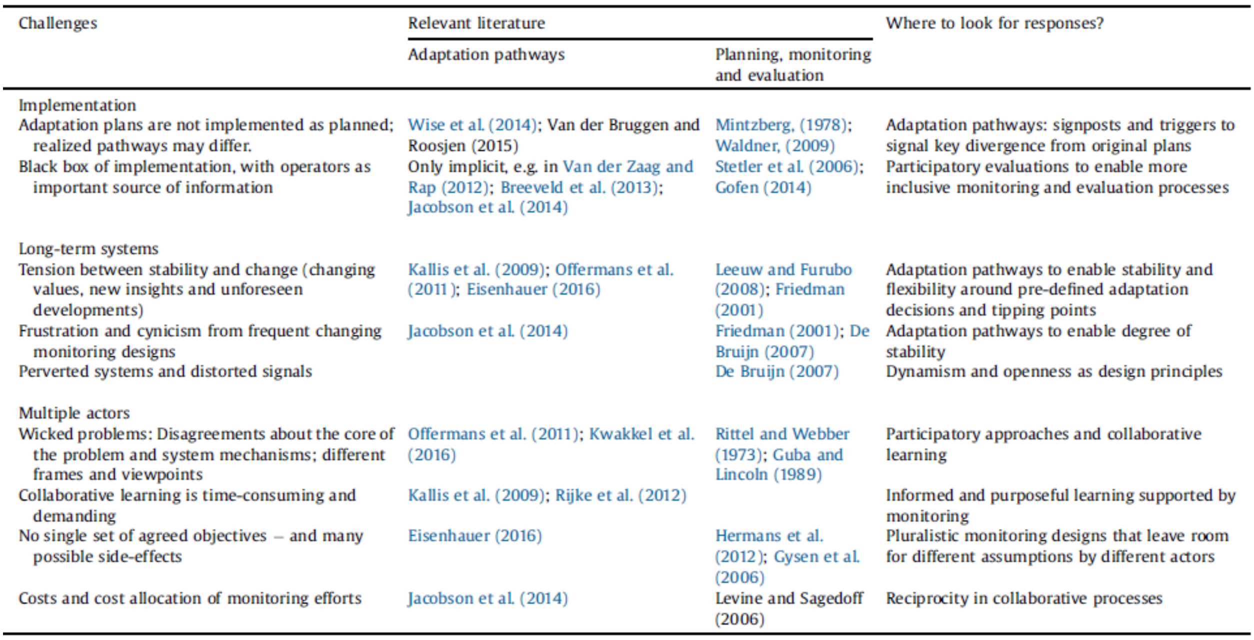
The call for monitoring and evaluation of adaptation pathways stems first and foremost from the expectation that adaptation pathways are being implemented and that the developed plans help identify the variables that need to be monitored (Hermans et al., 2017). However, Wise et al. (2014) suggest that adaptation plans are often not implemented and, if they are, it is only the smaller incremental measures within those plans. van der Brugge and Roosjen (2015) explain how the presumed implementation and effectiveness of adaptation pathways might be due to the changes in institutional and socio-cultural structures required for the implementation of adaptation strategies. This implementation problem is not unique to adaptation pathways (Hermans et al., 2017).

Some analysis indicates that a national and regional strategic approach, centered on a dynamic view of climate risk, is necessary for effective decisions at the local government and community level. In addition, effective adaptation requires better identification of barriers and opportunities to address changing risk, together with more effective and continuous social engagement (Manning et al., 2015). Moreover, a number of policy and legal shortcomings constrains integration and success in dynamic initiatives such as ICZM (Gerhartz-Abraham et al., 2016).

Adaptation pathways plans suggest policy actions for the short to medium term, within a longer-term pathway. These immediate policy actions are assembled in what is called a basic policy (Walker et al., 2013). In the pathways map shown in [Figure X], this basic policy corresponds to the “Current situation”, which provides the first path in an adaptation pathways plan. Monitoring may trigger a new decision, depicted as a decision node, which is to select and prepare for the appropriate ‘transfer station’ to a new policy action on the pathways map. Under different scenarios and for different time horizons, costs and benefits may be estimated for different sequences of policy actions. As simple as these analytic principles sound, various complications emerge for monitoring (Hermans et al., 2017).

Hermans et al. (2017) presents a key challenges in monitoring and evaluation of adaptation pathways and the responses to challenges, he concluded that, all in all, it is clear that collaborative learning about the implementation of adaptation pathways needs to be informed by monitoring and evaluation arrangements, and that there are many threats and challenges to its success. Table 4.10 summarizes these challenges, and suggests a first direction of where to look for responses.

Table 4.10: Challenges in the design of monitoring arrangements for adaptation pathways (Hermans et al., 2017, Table 1).



Challenges	Relevant literature		Where to look for responses?
	Adaptation pathways	Planning, monitoring and evaluation	
Implementation			
Adaptation plans are not implemented as planned; realized pathways may differ.	Wise et al. (2014); Van der Bruggen and Roosjen (2015)	Mintzberg, (1978); Waldner, (2009)	Adaptation pathways: signposts and triggers to signal key divergence from original plans
Black box of implementation, with operators as important source of information	Only implicit, e.g. in Van der Zaag and Rap (2012); Breeveld et al. (2013); Jacobson et al. (2014)	Stetler et al. (2006); Gofen (2014)	Participatory evaluations to enable more inclusive monitoring and evaluation processes
Long-term systems			
Tension between stability and change (changing values, new insights and unforeseen developments)	Kallis et al. (2009); Offermans et al. (2011); Eisenhauer (2016)	Leeuw and Furubo (2008); Friedman (2001)	Adaptation pathways to enable stability and flexibility around pre-defined adaptation decisions and tipping points
Frustration and cynicism from frequent changing monitoring designs	Jacobson et al. (2014)	Friedman (2001); De Bruijn (2007)	Adaptation pathways to enable degree of stability
Perverted systems and distorted signals		De Bruijn (2007)	Dynamism and openness as design principles
Multiple actors			
Wicked problems: Disagreements about the core of the problem and system mechanisms; different frames and viewpoints	Offermans et al. (2011); Kwakkel et al. (2016)	Rittel and Webber (1973); Guba and Lincoln (1989)	Participatory approaches and collaborative learning
Collaborative learning is time-consuming and demanding	Kallis et al. (2009); Rijke et al. (2012)		Informed and purposeful learning supported by monitoring
No single set of agreed objectives – and many possible side-effects	Eisenhauer (2016)	Hermans et al. (2012); Gysen et al. (2006)	Pluralistic monitoring designs that leave room for different assumptions by different actors
Costs and cost allocation of monitoring efforts	Jacobson et al. (2014)	Levine and Sagedoff (2006)	Reciprocity in collaborative processes

The challenges highlighted in Table 4.10 point to important trade-offs and some fundamental tensions, which cannot be resolved with a single specific procedure. However, they could think of approaches for the design of monitoring arrangements for adaptation pathways, which help the involved actors to find a workable balance for the trade-offs. They approached a propose for this has adaptation pathways as a central element, and is sketched in Figure 4.16. It consists of several building blocks, which explained in this article (Hermans et al., 2017).

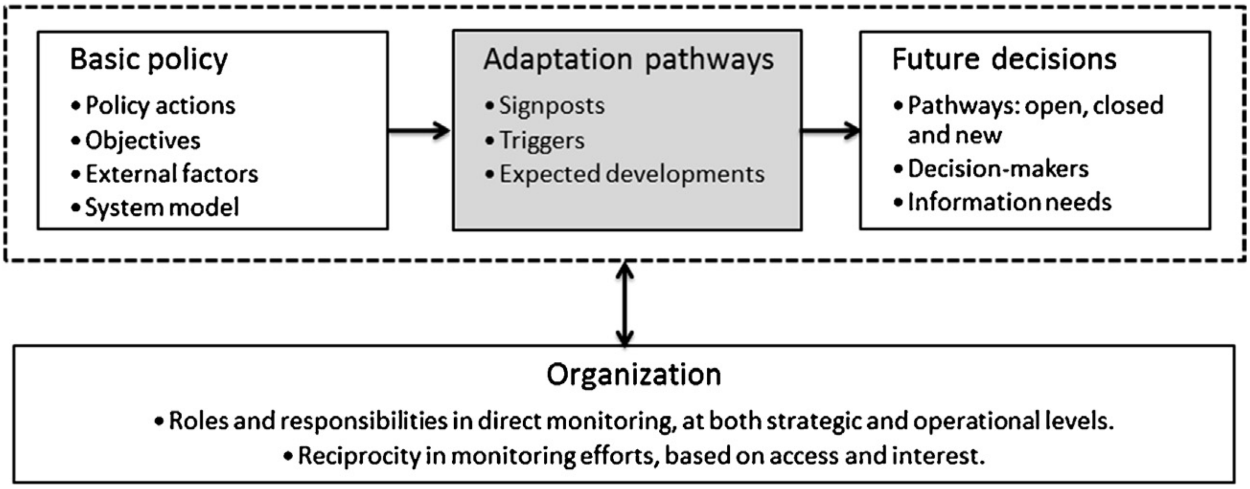


Figure 4.16: Approach for the Design of Monitoring Arrangements for Adaptation Pathways (Hermans et al., 2017, Figure 2).

4.4.7 Statement on Climate Resilient Development Pathways

[PLACEHOLDER FOR SECOND ORDER DRAFT: Sections 4.2, 4.3 and 4.4 to be completed and linked through medium of climate resilient development pathways. The focus on implications for low lying coastal cities, islands, deltas, social-ecological systems and communities, and the Arctic]

[START BOX 4.3 HERE]

Box 4.3: Case Studies on Coastal Hazard Risk and Rising Sea Levels

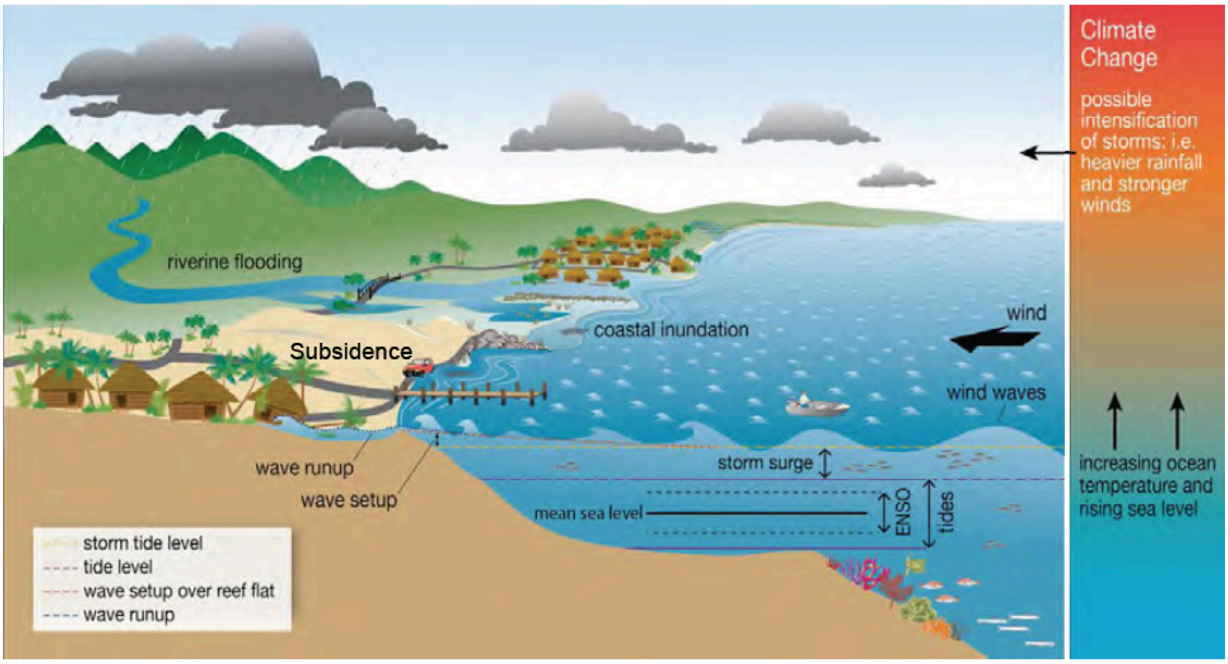
This box presents case studies that demonstrate how responses to coastal hazard risk and rising sea levels unfold in very different ways.

[PLACEHOLDER FOR SECOND ORDER DRAFT: case study on the Egyptian coast to be added; and a comparative case study analysis to be developed]

Coastal Flooding and Inundation, Nadi, Fiji

Source: Hay, J.E, 2017: Nadi Flood Control Project: Climate Risk and Vulnerability Assessment. Asian Development Bank, Manila, 60pp.

The Nadi River basin and Nadi Town, the third-largest conurbation in Fiji, are located on the western side of the main island of Viti Levu. Box 4.3, Figure 1, illustrates the main natural hazards that contribute to riverine flooding and coastal inundation for the Nadi Basin, namely heavy rainfall, elevated sea levels and subsidence of the delta. Tropical cyclones are particularly hazardous because of their potential to also elevate coastal sea levels due to storm surges and high waves. In addition to causing flooding of low-lying coastal terrain, higher coastal sea levels during a storm surge can slow the drainage of floodwater from coastal river systems to the ocean. This in turn may worsen the severity and extent of coastal and upstream flooding by a process that is referred to as the 'backwater' effect.



Box 4.3, Figure 1: Natural hazards that contribute to flooding and inundation. Adapted from McInnes (2014).

People and built assets in the Nadi River flood plain are already being affected by climate change. Observed sea level shows a long-term trend of 4 mm/year. But this is nearly obscured by interannual and other variability associated with tropical cyclone and El Niño events. Over the past 75 years the return periods of extreme rainfall events have decreased significantly. This is reflected in the fact that, of the 84 floods which have occurred in the Nadi River Basin since 1870, 54 occurred post 1980. There have been 26 major floods since 1991.

But the increased frequency of flooding is not all attributable to increases in sea level and extreme rainfall events. River channels have become filled with sediment, largely owing to deforestation of the hinterland. Much of the mangrove fringe has been sacrificed for development of various kinds. Like all river deltas, the one on which Nadi is located is subsiding.

The flood that occurred in March, 2012, is considered the largest historical flood on record, with a 50-year return period. It affected more than 150,000 people, with 4 deaths. In January, 2009 large areas of Fiji were inundated by devastating floods which claimed over 11 lives, left 12,000 people temporarily homeless and caused FJ$113 million of damage. Worst hit was the Nadi area, with total damage estimated at FJ$81.2 million.

Exceptionally high sea levels are associated with coastal inundation, accelerated coastal erosion and salt-water intrusion into groundwater. There is a high level of exposure to inundation for most of the Nadi flood plain, with the potential for a serious disaster if a 1-in-100 year design flood were to occur.

Projected changes in the frequency and intensity of tropical cyclones for mid- to late-21st century result in the more extreme sea levels (i.e., return periods of 200 years or more) becoming higher, while those with return periods of 50 years and less become lower. For 1-in-100 year events heights were little changed. Overall, projected changes in sea level were found to make the largest contribution to increased extreme sea level risk (McInnes and Hoeke, 2014; McInnes et al., 2014).

Various initiatives to help alleviate flooding and inundation in the Nadi basin have been proposed. These include both structural (e.g., ring dikes, river widening, bridge rebuilding, retarding basins, shortcutting tributaries, dams and diversion channels) and non-structural (e.g., early flood warnings, improved land management practices in upper basin) interventions.

Box 4.3, Table 1 shows that recent improvements in understanding call for significant changes in the basis for the design and planning of the structural and related interventions. The baseline is the understanding that existed prior to that which is now captured in SROCC.

Box 4.3, Table 1: Changes in the Basis for Design and Planning of Structural and Related Interventions

	Baseline Assessment	Consistent with SROCC Assessment
Hazards	Design storm tide: 1.2 m 2-day design rainfall: 436mm Subsidence: Not considered	Design storm tide: 2.31m (100-year extreme sea level in 2055) Two day design rainfall: 670 mm Subsidence: 0.4mm/year
Exposure and Vulnerability	Exposure and vulnerability assessed for present day only—thus static, with no reference to drivers	Exposure and vulnerability assessed for the present day and future time periods, with the projections taking into account both bio-physical and human drivers
Levels of Risk	Reflect current levels of risk, with no allowance for climate, biophysical or socio-economic changes	Risks reflect the full suite of biophysical and socio-economic changes over the life of the planned investment project, including their interactions
Response Options	Interventions based entirely on reducing current levels of risk, with the primary focus being on structural measures to reduce flood hazard, and thereby flood risk. Non-structural measures not prioritised	Rational mix of structural and non-structural interventions to reduced risks likely to occur over and beyond the life of the planned investment project
Planning and Decision Making	Takes a narrow 'flood control' approach aimed at 'controlling' single hazards, rather than managing the multiple and interacting risks in their broader contexts.	Takes a risk-based, flexible design approach that addresses the tension between the constancy of a given design standard on the one hand and, on the other hand, increasing flood risk over time due to further floodplain development, climate change leading to higher peak flows and inundation, and river channel bed aggradation

A comparison of coastal flood hazard, vulnerability and adaptation measures between New York City and Shanghai

New York City (NYC) is the financial center of the US and lies at the junction of the Hudson River and Atlantic Ocean. Shanghai is located at the mouth of the Yangtze River where it enters the East China Sea and it is the economic center of China. Both cities play critical roles in the global economy and trade, with dense population, infrastructure, and concentrated assets in the floodplain (e.g., Lower Manhattan in NYC) and a long history of extreme flooding events.

Hurricane Sandy (2012) and Typhoon Winnie (1997) are considered to be the largest historical flood events for NYC and Shanghai, respectively. Hurricane Sandy killed 55 people in the US and caused over USD32 billion losses for US. New Jersey and New York areas witnessed the most substantial damage along the coastlines (Xian et al., 2015). Typhoon Winnie killed more than 310 people and caused damage exceeding USD3.2 billion to China. Many dikes and flood walls along coastal Shanghai and Zhejiang were breached by surge flood waters. Storm surge and heavy rainfall inundated many parts of the towns and cities as the typhoon moved inland. Previous studies estimate the return period of a flood reaching the levels attained during Hurricane Sandy at from 100 to 1200 years at the Battery tide gauge (Sweet et al., 2013; Zervas, 2013; Lopeman et al., 2015; Lin et al., 2016; Xian et al., 2018). The return period of the flood level attained during Typhoon Winnie is about 100 years at the Wusong tide gauge station (Yin et al., 2013; Xian et al., 2018).

The two cities face an increasing flood risks in the future due to sea level rise and local land subsidence. NYC's sea level at the Battery station rose at an average rate of 1.3 mm yr^{-1} (excluding land subsidence) over the last 100 years (Xian et al., 2018). Sea level at the Wusong station in Shanghai rose at a faster average rate of 2.6 mm yr^{-1} (again excluding land subsidence) over the period of 1910–2000 (Yin et al., 2011). Land subsidence in Shanghai, estimated at 5mm/year, is also much higher than at NYC. The Shanghai rate is dominated by tectonic subsidence (TS) and compaction of sediments (Shanghai Municipal Bureau of Planning and Land Resources, 2007; Gong and Yang, 2008; Yin et al., 2013). Therefore, the relative sea level rise in Shanghai is considerably higher than NYC. Both rates far exceed the global mean rate of rise over 20th century.

The economic exposure of both cities is also high. Hanson et al. (2011) estimated that for 2005 the value of exposed coastal assets in the New York-Newark area (USD 320 billion) was over four times that of Shanghai (USD 73 billion). By 2070 the magnitude of the exposed assets of Shanghai (USD 1.7 trillion) is expected to be close to that of the New York-Newark metropolitan area (USD 2.1 trillion; Hanson et al., 2011). Limited construction and development activities have occurred in NYC (especially in Manhattan, Brooklyn, and the Bronx) since 1979, compared with the rapid development in Shanghai, whose urban area increased by 1064% from 1979 to 2010 owing to the rapid economic transition and development in China (Xian et al., 2018). Most of Shanghai’s growth in exposed value has occurred near the low elevation urban center. Future development along the coast is *likely* to be greater for Shanghai than NYC.

In addition, individual past extreme flood events may have influenced the immediate reaction of policy makers and can be the driving force for protection measures (Pelling and Dill, 2010; Albright, 2011). If we overlay the past updates of the flood defense heights on top of the past annual maximum water levels at the Wusong station in Shanghai, we found that each update is associated with an extreme flood event induced by a severe typhoon (i.e., typhoons in 1962, 1974, and 1981). In contrast, the peak water tide of Hurricane Sandy stands out in the record at the Battery tidal gauge station. Unlike frequent attacks by severe flood inundation, NYC suffered relatively moderate consequences from individual events before Hurricane Sandy. All these factors together induced higher-standard flood protection measures in Shanghai, such as sea walls with a 200-year coastal flood return level design that protect its coastlines and critical infrastructure of the developed areas, and flood walls with 1000-year riverine flood return level along the Huangpu River to protect the city from riverine flooding. New York City, on the other hand, has relatively lower protection, consisting of sandy dunes (e.g., on Staten Island), vegetation (e.g., in Queens) and low-rise sea walls in lower (Manhattan). Since Hurricane Sandy in 2012, discussions about possible flood protection strategies for concentrated assets and infrastructure engaged a range of stakeholder groups. For example, implementation of the ‘Big U’ project, a proposed coastal protection system for lower Manhattan, has begun and new public and private hospitals are required to be out of the flood zones. Moreover, the MTA introduced a new equipment that includes custom doors and curtains that can be deployed to protect underground subway stations and can withstand 4.3 m of water above street level from future flooding.

In spite of higher-standard of flood protection measures, the current protections in Shanghai may also not be sufficient to prevent future flooding. Previous studies show that around half of the length of current sea walls in Shanghai may be overtopped by storms in 2100 (Wang et al., 2012). For any coastal megacities just like Shanghai and NYC, if policy makers are reluctant to spend on protection until disaster strikes, the result can be more costly actions later, e.g., demolishing and then reconstruction of sea walls, from the past experience of Shanghai (Xian et al., 2018). A better way may be to incorporate the time-variant hazard systematically into the design process (Lickley et al., 2014). Such an alternative, flexible approach would be to update defense every a few years of interval to adjust for the new climate risk information and ensure that the protection height is above the level of the acceptable risk.

Governance and funding structure are also important to consider when interpreting effectiveness of adaptation by megacities. For example, Shanghai has a high level of autonomy in decision making, and objectives of the central and local governments are aligned, allowing them to achieve consensus on extensive efforts needed to protect the city (Wei and Leung, 2005; Yin et al., 2011). In addition, rapid economic growth in Shanghai and China during the past 30 years provided adequate funding for large-scale infrastructure (Zhang, 2003). In contrast, the multiple jurisdictions of the city, the state and federal government in the case of NYC present a challenge to implementing measures effectively to mitigate climate-related risks (Rosenzweig and Solecki, 2014). NYC may learn from the experience of London (Thames Barrier), Rotterdam, Amsterdam (dike-rings) and New Orleans (surge barrier) that are all good examples in democratic states and have invested significantly in large flood protection infrastructure [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to be included: Sayvetz, 2015].

Box 4.3, Table 2: Changes in the Basis for Design and Planning of Structural and Related Interventions

	Current Situation	Reflecting SROCC Assessment
Hazards Assessment	100-yr storm tide: 5.87 m Shanghai vs. 3.29 m NYC. Subsidence: 5 mm yr^{-1} Shanghai vs 1-2mm yr^{-1} NYC.	Design storm tide: 6.1^{+} m Shanghai vs 3.5^{+} m NYC (100-year flood by 2050). Subsidence in Shanghai may be higher due to rapid population increase.
Multiple Drivers of Exposure and Vulnerability	Current exposure & vulnerability: topographic ground elevation, social vulnerability, economic vulnerability (assets in low elevation)	Exposure and vulnerability assessed for the future time periods: population change, infrastructure planning, urban expansion, asset increase
Levels of Risk	Reflect current levels of risk plus some levels of freeboard:0.3–1 m for NYC; 0.5^{+} m for Shanghai, with no consideration for storm characteristic and socio-economic changes	Risks reflect the full suite of climate change and socio-economic changes for the planned coastal projects and their interdependency; use the 90th percentile sea level as the freeboard criteria to enhance the safety of protection
Response Options	Structural measures to reduce flood hazard, including fixed height sea wall, building retrofit; Non-structural: catastrophe insurance and natural defenses	Optimal mix of structural and non-structural measures to reduced flood risks to minimize the combination of coast and expected future losses; more flexible structural measures that reflect the future changing risks
Planning and Decision Making	Funding and governance issues that may slow down the planning and implementation process; the long-term uncertainty of long-term risk estimation may restrict current actions on flood protection	Consider the adaptation of policy responses onto the long-term risk might reduce the fear towards long-term protection planning and trigger a more adaptive and flexible policy measure

[END BOX 4.3 HERE]

[START FAQ4.1 HERE]

FAQ4.1: What makes communities especially vulnerable to coastal hazards related to climate change?

[END FAQ4.1 HERE

[START FAQ4.2 HERE]

FAQ4.2: What challenges do coastal communities face when adapting to projected sea level rise?

[END FAQ4.2 HERE

[START FAQ4.3 HERE]

FAQ4.3: Are islands drowning because of climate change?

[END FAQ4.3 HERE

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