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Extremes, Abrupt Changes and Managing Risks

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Executive Summary

This chapter assesses extremes and abrupt or irreversible changes in the ocean and cryosphere in a changing climate, to identify regional hot spots, cascading effects, their impacts on human and natural systems, and sustainable and resilient risk management strategies. It is not comprehensive in terms of the systems assessed and some information on extremes, abrupt and irreversible changes, in particular for the cryosphere, may be found in other chapters.

Ongoing and Emerging Changes in the Ocean and Cryosphere, and their Impacts on Ecosystems and Human Societies

Anthropogenic climate change has increased observed precipitation (*medium confidence*), winds (*low confidence*), and extreme sea level events (*high confidence*) associated with some tropical cyclones, which has increased intensity of multiple extreme events and associated cascading impacts (*high confidence*). Anthropogenic climate change may have contributed to a poleward migration of maximum tropical cyclone intensity in the western North Pacific in recent decades related to anthropogenically-forced tropical expansion (*low confidence*). There is emerging evidence for an increase in the annual global proportion of Category 4 or 5 tropical cyclones in recent decades (*low confidence*). {6.3, Table 6.2, Figure 6.2, Box 6.1}

Changes in Arctic sea ice have the potential to influence mid-latitude weather (*medium confidence*), but there is *low confidence* in the detection of this influence for specific weather types. {6.3}

Extreme wave heights, which contribute to extreme sea level events, coastal erosion and flooding, have increased in the Southern and North Atlantic Oceans by around 1.0 cm yr^{-1} and 0.8 cm yr^{-1} over the period 1985–2018 (*medium confidence*). Sea ice loss in the Arctic has also increased wave heights over the period 1992–2014 (*medium confidence*). {6.3}

Marine heatwaves (MHWs), periods of extremely high ocean temperatures, have negatively impacted marine organisms and ecosystems in all ocean basins over the last two decades, including critical foundation species such as corals, seagrasses and kelps (*very high confidence*). Globally, marine heat related events have increased; marine heatwaves, defined when the daily sea surface temperature exceeds the local 99th percentile over the period 1982 to 2016, have doubled in frequency and have become longer-lasting, more intense and more extensive (*very likely*). It is *very likely* that between 84–90% of marine heatwaves that occurred between 2006 and 2015 are attributable to the anthropogenic temperature increase. {6.4, Figures 6.3, 6.4}

Both palaeoclimate and modern observations suggest that the strongest El Niño and La Niña events since the pre-industrial period have occurred during the last fifty years (*medium confidence*). There have been three occurrences of extreme El Niño events during the modern observational period (1982–1983, 1997–1998, 2015–2016), all characterised by pronounced rainfall in the normally dry equatorial East Pacific. There have been two occurrences of extreme La Niña (1988–1989, 1998–1999). El Niño and La Niña variability during the last 50 years is unusually high compared with average variability during the last millennium. {6.5, Figure 6.5}

The equatorial Pacific trade wind system experienced an unprecedented intensification during 2001–2014, resulting in enhanced ocean heat transport from the Pacific to the Indian Ocean, influencing the rate of global temperature change (*medium confidence*). In the last two decades, total water transport from the Pacific to the Indian Ocean by the Indonesian Throughflow (ITF), and the Indian Ocean to Atlantic Ocean has increased (*high confidence*). Increased ITF has been linked to Pacific cooling trends and basin-wide warming trends in the Indian Ocean. Pacific sea surface temperature (SST) cooling trends and strengthened trade winds have been linked to an anomalously warm tropical Atlantic. {6.6, Figure 6.7}

Observations, both in situ (2004–2017) and based on sea surface temperature reconstructions, indicate that the Atlantic Meridional Overturning Circulation (AMOC) has weakened relative to 1850–1900 (*medium confidence*). There is insufficient data to quantify the magnitude of the weakening, or to properly attribute it to anthropogenic forcing due to the limited length of the observational record. Although attribution is currently not possible, CMIP5 model simulations of the period 1850–2015, on average, exhibit a weakening AMOC when driven by anthropogenic forcing. {6.7, Figure 6.8}

Climate change is modifying multiple types of climate-related events or hazards in terms of occurrence, intensity and periodicity. It increases the likelihood of compound hazards that comprise simultaneously or sequentially occurring events to cause extreme impacts in natural and human systems. Compound events in turn trigger cascading impacts (*high confidence*). Three case studies are presented in the chapter, (i) Tasmania's Summer of 2015–2016, (ii) The Coral Triangle and (ii) Hurricanes of 2017. {6.8, Box 6.1}

Projections of Ocean and Cryosphere Change and Hazards to Ecosystems and Human Society Under Low and High Emission Futures

The average intensity of tropical cyclones, the proportion of Category 4 and 5 tropical cyclones and the associated average precipitation rates are projected to increase for a 2°C global temperature rise above any baseline period (*medium confidence*). Rising mean sea levels will contribute to higher extreme sea levels associated with tropical cyclones (*very high confidence*). Coastal hazards will be exacerbated by an increase in the average intensity, magnitude of storm surge and precipitation rates of tropical cyclones. There are greater increases projected under RCP8.5 than under RCP2.6 from around mid-century to 2100 (*medium confidence*). There is *low confidence* in changes in the future frequency of tropical cyclones at the global scale. {6.3.1}

Significant wave heights (the average height from trough to crest of the highest one-third of waves) are projected to increase across the Southern Ocean and tropical eastern Pacific (*high confidence*) and Baltic Sea (*medium confidence*) and decrease over the North Atlantic and Mediterranean Sea under RCP8.5 (*high confidence*). Coastal tidal amplitudes and patterns are projected to change due to sea level rise and coastal adaptation measures (*very likely*). Projected changes in waves arising from changes in weather patterns, and changes in tides due to sea level rise, can locally enhance or ameliorate coastal hazards (*medium confidence*). {6.3.1, 5.2.2}

Marine heatwaves are projected to further increase in frequency, duration, spatial extent and intensity (maximum temperature) (*very high confidence*). Climate models project increases in the frequency of marine heatwaves by 2081–2100, relative to 1850–1900, by approximately 50 times under RCP8.5 and 20 times under RCP2.6 (*medium confidence*). The largest increases in frequency are projected for the Arctic and the tropical oceans (*medium confidence*). The intensity of marine heatwaves is projected to increase about 10-fold under RCP8.5 by 2081–2100, relative to 1850–1900 (*medium confidence*). {6.4}

Extreme El Niño and La Niña events are projected to *likely* increase in frequency in the 21st century and to *likely* intensify existing hazards, with drier or wetter responses in several regions across the globe. Extreme El Niño events are projected to occur about as twice as often under both RCP2.6 and RCP8.5 in the 21st century when compared to the 20th century (*medium confidence*). Projections indicate that extreme Indian Ocean Dipole events also increase in frequency (*low confidence*). {6.5; Figures 6.5, 6.6}

Lack of long-term sustained Indian and Pacific Ocean observations, and inadequacies in the ability of climate models to simulate the magnitude of trade wind decadal variability and the inter-ocean link, mean there is *low confidence* in future projections of the trade wind system. {6.6, Figure 6.7}

The AMOC will *very likely* weaken over the 21st century (*high confidence*), although a collapse is *very unlikely* (*medium confidence*). Nevertheless, a substantial weakening of the AMOC remains a physically plausible scenario. Such a weakening would strongly impact natural and human systems, leading to a decrease in marine productivity in the North Atlantic, more winter storms in Europe, a reduction in Sahelian and South Asian summer rainfall, a decrease in the number of TCs in the Atlantic, and an increase in regional sea level around the Atlantic especially along the northeast coast of North America (*medium confidence*). Such impacts would be superimposed on the global warming signal. {6.7, Figure 6.8}

Impacts from further changes in TCs and ETCs, MHWs, extreme El Niño and La Niña events and other extremes will exceed the limits of resilience and adaptation of ecosystems and people, leading to unavoidable loss and damage (*medium confidence*). {6.9.2}

Strengthening the Global Responses in the Context of Sustainable Development Goals (SDGs) and Charting Climate Resilient Development Pathways for Oceans and Cryosphere

There is *medium confidence* that including extremes and abrupt changes, such as AMOC weakening, ice sheet collapse (West Antarctic Ice Sheet (WAIS) and Greenland Ice Sheet (GIS)), leads to a several-fold increase in the cost of carbon emissions (*medium confidence*). If carbon emissions decline, the risk of extremes and abrupt changes are reduced, creating co-benefits. {6.8.6}

For TCs and ETCs, investment in disaster risk reduction, flood management (ecosystem and engineered) and early warning systems decreases economic loss (*medium confidence*), but such investments may be hindered by limited local capacities, such as increased losses and mortality from extreme winds and storm surges in less developed countries despite adaptation efforts. There is emerging evidence of increasing risks for locations impacted by unprecedented storm trajectories (*low confidence*). Managing the risk from such changing storm trajectories and intensity proves challenging because of the difficulties of early warning and its receptivity by the affected population (*high confidence*). {6.3, 6.9}

Limiting global warming would reduce the risk of impacts of MHWs, but critical thresholds for some ecosystems (e.g., kelp forests, coral reefs) will be reached at relatively low levels of future global warming (*high confidence*). Early warning systems, producing skillful forecasts of MHWs, can further help to reduce the vulnerability in the areas of fisheries, tourism and conservation, but are yet unproven at large scale (*medium confidence*). {6.4}

Sustained long-term monitoring and improved forecasts can be used in managing the risks of extreme El Niño and La Niña events associated with human health, agriculture, fisheries, coral reefs, aquaculture, wildfire, drought and flood management (*high confidence*). {6.5}

Extreme change in the trade wind system and its impacts on global variability, biogeochemistry, ecosystems and society have not been adequately understood and represent significant knowledge gaps. {6.6}

By 2300, an AMOC collapse is *as likely as not* for high emission pathways and *very unlikely* for lower ones, highlighting that an AMOC collapse can be avoided in the long term by CO_2 mitigation (*medium confidence*). Nevertheless, the human impact of these physical changes have not been sufficiently quantified and there are considerable knowledge gaps in adaptation responses to a substantial AMOC weakening. {6.7}

The ratio between risk reduction investment and reduction of damages of extreme events varies. Investing in preparation and prevention against the impacts from extreme events is *very likely* less than the cost of impacts and recovery (*medium confidence*). Coupling insurance mechanisms with risk reduction measures can enhance the cost-effectiveness of adapting to climate change (*medium confidence*). {6.9}

Climate change adaptation and disaster risk reduction require capacity building and an integrated approach to ensure trade-offs between short- and long-term gains in dealing with the uncertainty of increasing extreme events, abrupt changes and cascading impacts at different geographic scales (*high confidence*). {6.9}

Limiting the risk from the impact of extreme events and abrupt changes leads to successful adaptation to climate change with the presence of well-coordinated climate-affected sectors and disaster management relevant agencies (*high confidence*). Transformative governance inclusive of successful integration of disaster risk management (DRM) and climate change adaptation, empowerment of vulnerable groups, and accountability of governmental decisions promotes climate-resilient development pathways (*high confidence*). {6.9}

6.1 Introduction

This chapter assesses extremes and abrupt or irreversible changes in the ocean and cryosphere in a changing climate, to identify regional hot spots, cascading effects, their impacts on human and natural systems, and sustainable and resilient risk management strategies. While not comprehensive in terms of discussing all such phenomena, it addresses a number of issues that are prominent in both the policy area and in the scientific literature. Further information may also be found in Chapters 2 to 4 for other aspects of the ocean and cryosphere.

Building on the Special Report on Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation (SREX; IPCC, 2012), IPCC 5th Assessment Report (AR5; IPCC, 2013; IPCC, 2014) assessments and the Special Report on Global Warming of 1.5°C (SR15; IPCC, 2018), for each of the topics addressed, we provide an assessment of:

- Key processes and feedbacks, observations, detection and attribution, projections;
- Impacts on human and natural systems;
- Monitoring and early warning systems;
- Risk management and adaptation, sustainable and resilient pathways.

The chapter is organised in terms of the space- and time-scales of different phenomena. We move from small-scale TCs, which last for days to weeks, to the global-scale AMOC, which has time scales of decades to centuries. A common risk framework is adopted, based on that used in AR5 and introduced in Chapter 1, Section 1.5 and Cross-Chapter Box 1 in Chapter 1 (Figure 6.1).

While much of what is discussed within the chapter concerns the ocean, we also summarise abrupt events in the cryosphere in Section 6.2, drawing information from Chapters 2 to 4, where the main assessment of those phenomena may be found.

6.1.1 Definitions of Principal Terms

In discussing concepts such as abrupt changes, irreversibility, tipping points and extreme events it is important to define precisely what is meant by those terms. The following definitions are therefore adopted (based on either AR5, Special Report on Global Warming of 1.5°C (SR15) or Special Report on Climate Change and Land (SRCCL) Glossaries):

Abrupt climate change: A large-scale change in the climate system that takes place over a few decades or less, persists (or is anticipated to persist) for at least a few decades, and causes substantial disruptions in human and natural systems.

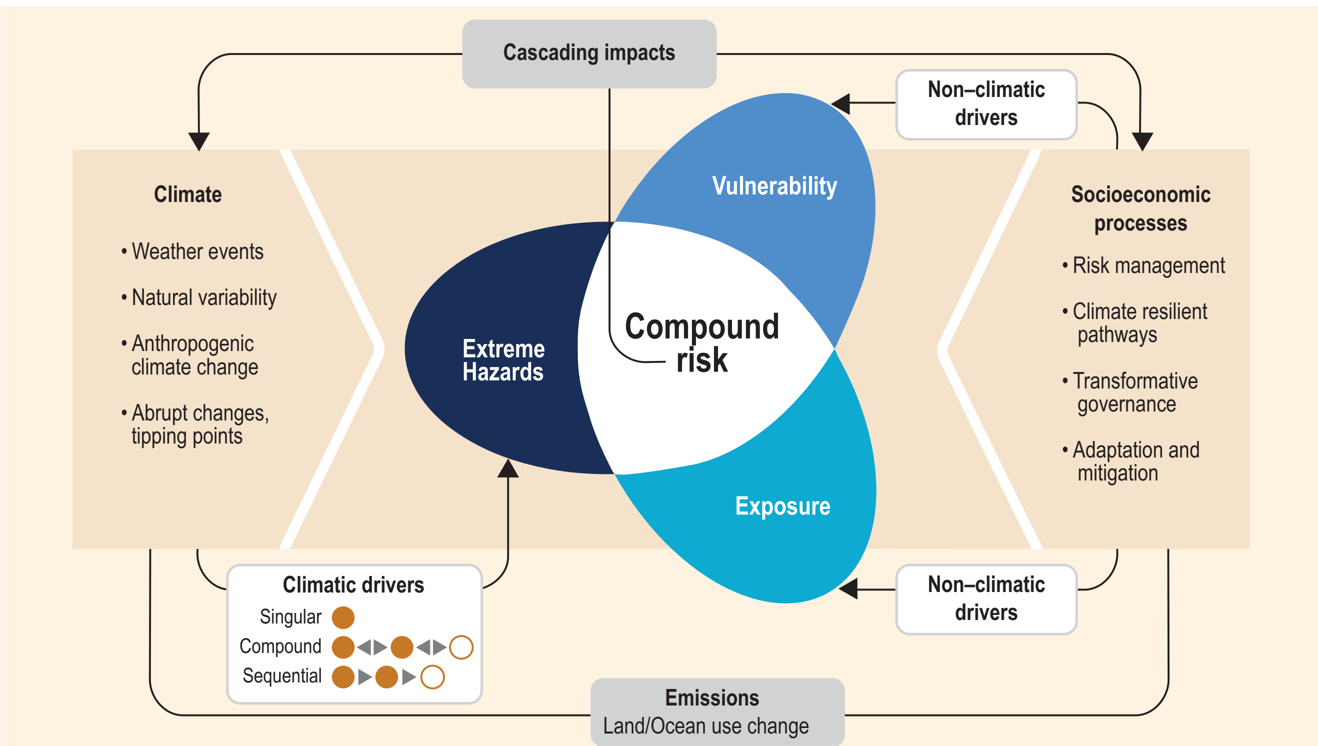

Figure 6.1 | Framework used in this chapter (see discussion in Chapter 1). Singular or multiple climate drivers can lead to extreme hazards and associated cascading impacts, which combined with non-climatic drivers affect exposure and vulnerability, leading to compound risks. Extremes discussed are tropical cyclones (TCs) and extratropical cyclones (ETCs) and associated sea surface dynamics (Section 6.3); marine heatwaves (MHWs) (Section 6.4), extreme El Niño and La Niña events (Section 6.5); and extreme oceanic decadal variability (Section 6.6). Examples of abrupt events, irreversibility and tipping points discussed are the Atlantic Meridional Overturning Circulation (AMOC) and subpolar gyre (SPG) system (Section 6.7). Section 6.2 also collects examples of such events from the rest of the Special Report on the Oceans and Cryosphere in a Changing Climate (SROCC) and compiles examples of events whose occurrence or severity has been linked to climate change. Cascading impacts and compound events are discussed in Section 6.8 and three examples are given in Box 6.1. Section 6.9 discusses risk management, climate resilience pathways, transformative governance adaptation and mitigation required to address societal and environmental risks.

Extreme weather/climate event: An extreme event is an event that is rare at a particular place and time of year. Definitions of 'rare' vary, but an extreme event would normally be as rare as or rarer than the 10th or 90th percentile of a probability density function estimated from observations. By definition, the characteristics of what is called an extreme event may vary from place to place in an absolute sense. When a pattern of extreme weather persists for some time, such as a season, it may be classed as an extreme climate event, especially if it yields an average or total that is itself extreme (e.g., high temperature, drought, or total rainfall over a season).

Irreversibility: A perturbed state of a dynamical system is defined as irreversible on a given timescale, if the recovery timescale from this state due to natural processes is significantly longer than the time it takes for the system to reach this perturbed state. In the context of this report, the recovery time scale of interest is hundreds to thousands of years.

Tipping point: A level of change in system properties beyond which a system reorganises, often in a nonlinear manner, and does not return to the initial state even if the drivers of the change are abated. For the climate system, the term refers to a critical threshold when global or regional climate changes from one stable state to another stable state. Tipping points are also used when referring to impact; the term can imply that an impact tipping point is (about to be) reached in a natural or human system.

These above four terms generally refer to aspects of the physical climate system. Here we extend their definitions to natural and human systems. For example, there may be gradual physical climate change which causes an irreversible change in an ecosystem. An adaptation tipping point could be reached when an adaptation option no longer remains effective. There may be a tipping point within a governance structure.

We also introduce two new key terms relevant for discussing risk-related concepts:

Compound events refer to the combination of multiple drivers and/or hazards that contribute to societal or environmental risks.

Cascading impacts from extreme weather/climate events occur when an extreme hazard generates a sequence of secondary events in natural and human systems that result in physical, natural, social or economic disruption, whereby the resulting impact is significantly larger than the initial impact. Cascading impacts are complex and multi-dimensional, and are associated more with the magnitude of vulnerability than with that of the hazard.

6.2 Climate Change influences on Abrupt Changes, Irreversibility, Tipping Points and Extreme Events

6.2.1 Introduction

Some potentially abrupt or irreversible events are assessed in other chapters, hence Table 6.1 presents a cross-chapter summary of those. Subsection numbers indicate where detailed information may be found.

Table 6.1 | Cross-Chapter assessment of abrupt and irreversible phenomena related to the ocean and cryosphere. The column on the far right of the table indicates the likelihood of an abrupt/irreversible change based on the assessed literature which, in general, assesses Representative Concentration Pathway (RCP) scenarios. Assessments of likelihood and confidence are made according to IPCC guidance on uncertainties.

Change in system component	Potentially abrupt	Irreversibility if forcing reversed (time scales indicated)	Impacts on natural and human systems; global *vs.* regional *vs.* local	Projected likelihood and/or confidence level in 21st century under scenarios considered
Ocean				
Atlantic Meridional Overturning Circulation (AMOC) collapse (Section 6.7)	Yes	Unknown	Widespread; increased winter storms in Europe, reduced Sahelian rainfall and agricultural capacity, variations in tropical storms, increased sea levels on Atlantic coasts	*Very unlikely*, but physically plausible
Subpolar gyre (SPG) cooling (Section 6.7)	Yes	Irreversible within decades	Similar to AMOC impacts but considerably smaller	*Medium confidence*
Marine heatwave (MHW) increase (Section 6.4)	Yes	Reversible within decades to centuries	Coral bleaching, loss of biodiversity and ecosystem services, harmful algal blooms, species redistribution	*Very likely* (*very high confidence*) for physical change *High confidence* for impacts
Arctic sea ice retreat (Section 3.3)	Yes	Reversible within decades to centuries	Coastal erosion in Arctic (may take longer to reverse), impact on mid-latitude storms (*low confidence*); rise in Arctic surface temperatures (*high confidence*)	*High confidence*
Ocean deoxygenation and hypoxic events (Section 5.2)	Yes	Reversible at surface, but irreversible for centuries to millennia at depth	Major changes in ocean productivity, biodiversity and biogeochemical cycles	*Medium confidence*
Ocean acidification (Section 5.2)	Yes	Reversible at surface, but irreversible for centuries to millennia at depth	Changes in growth, development, calcification, survival and abundance of species, for example, from algae to fish	*Virtually certain* (*very high confidence*)

Change in system component	Potentially abrupt	Irreversibility if forcing reversed (time scales indicated)	Impacts on natural and human systems; global *vs.* regional *vs.* local	Projected likelihood and/or confidence level in 21st century under scenarios considered
Cryosphere				
Methane release from permafrost (Section 3.4)	Yes	Reversible due to short lifetime of methane in the atmosphere	Further increased global temperatures through climate feedback	*Medium confidence*
CO_2 release from permafrost (Section 3.4)	Yes	Irreversible for millennia due to long lifetime of CO_2 in the atmosphere	Further increased global temperatures through climate feedback	*Low confidence*
Partial West Antarctic Ice Sheet (WAIS) collapse (Cross Chapter Box 2 in Chapter 1, Section 4.2)	Yes (late 21st century, under RCP8.5 only)	Irreversible for decades to millennia	Significant contribution to sea level rise (SLR) and local decrease in ocean salinity	*Low confidence*
Greenland Ice sheet (GIS) decay (Cross Chapter Box 8, Section 4.2)	No	Irreversible for millennia	Significant contribution to SLR, shipping (icebergs)	*High confidence* for decay contributing 10s of cm of SLR
Ice-shelf collapses (Cross Chapter Box 8, Sections 3.3, 4.2)	Yes	Possibly irreversible for centuries	May lead to SLR from contributing glaciers; some shelves more prone than others	*Low confidence*
Glacier avalanches, surges, and collapses (Section 2.3)	Yes	Variable	Local hazard; may accelerate SLR; local iceberg production; local ecosystems	*Medium confidence* for occurrence *Low confidence* for increase in frequency/magnitude
Strong shrinkage or disappearance of individual glaciers (Sections 2.2, 3.3)	Yes	Reversible within decades to centuries	Regional impact on water resources, tourism, ecosystems and global sea level	*Medium confidence*
Landslides related to glaciers and permafrost, glacier lake outbursts (Section 2.3)	Yes	Irreversible for rock slopes; reversible within decades to centuries for glaciers, debris and lakes	Local direct impact on humans, land use, infrastructure (hazard), and ecosystems	*Medium confidence for increase in frequency*
Change in biodiversity in high mountain areas (impact – Section 2.4)	Yes	In many cases irreversible (e.g., extinction of species)	Local impacts on ecosystems and ecosystem services	*Medium confidence*

6.2.2 Recent Anomalous Extreme Climate Events and their Causes

The attribution of changes in the observed statistics of extremes are generally addressed using well-established detection-attribution methods. In contrast, record-breaking weather and climate events are by definition unique, and can be expected to occur with or without climate change as the observed record lengthens. Therefore, event attribution begins with the premise that the climate is changing, the goal being to determine statistically how much climate change has contributed to the severity of the event (Trenberth et al. 2015; Shepherd, 2016). Annual reports dedicated to extreme event attribution (Peterson et al. 2012; Peterson et al. 2013; Herring et al. 2014; Herring et al. 2015; Herring et al. 2018) have helped stimulate studies that adopt recognised methods for extreme event attribution. The increasing pool of studies allows different approaches to be contrasted and builds consensus on the role of climate change when individual climate events are studied by multiple teams using different methods. A number of these events are summarised in Table 6.2 and Figure 6.2. Collectively, these studies show that the role of climate change in the ocean and cryosphere extreme events is increasingly driving extreme climate and weather events across the globe including compound events (*high confidence*). Some regions including Africa and the Pacific have had relatively fewer event attribution studies undertaken, possibly reflecting the lack of capacity by regional and national technical institutions. A caveat of this approach is that there is a potential for 'null results', that is, cases where attribution is not possible, to be reported. Nevertheless, there is no evidence that this is the case, and the number of recent studies and wide range of phenomena addressed suggests increasing influence of climate change on extreme events.

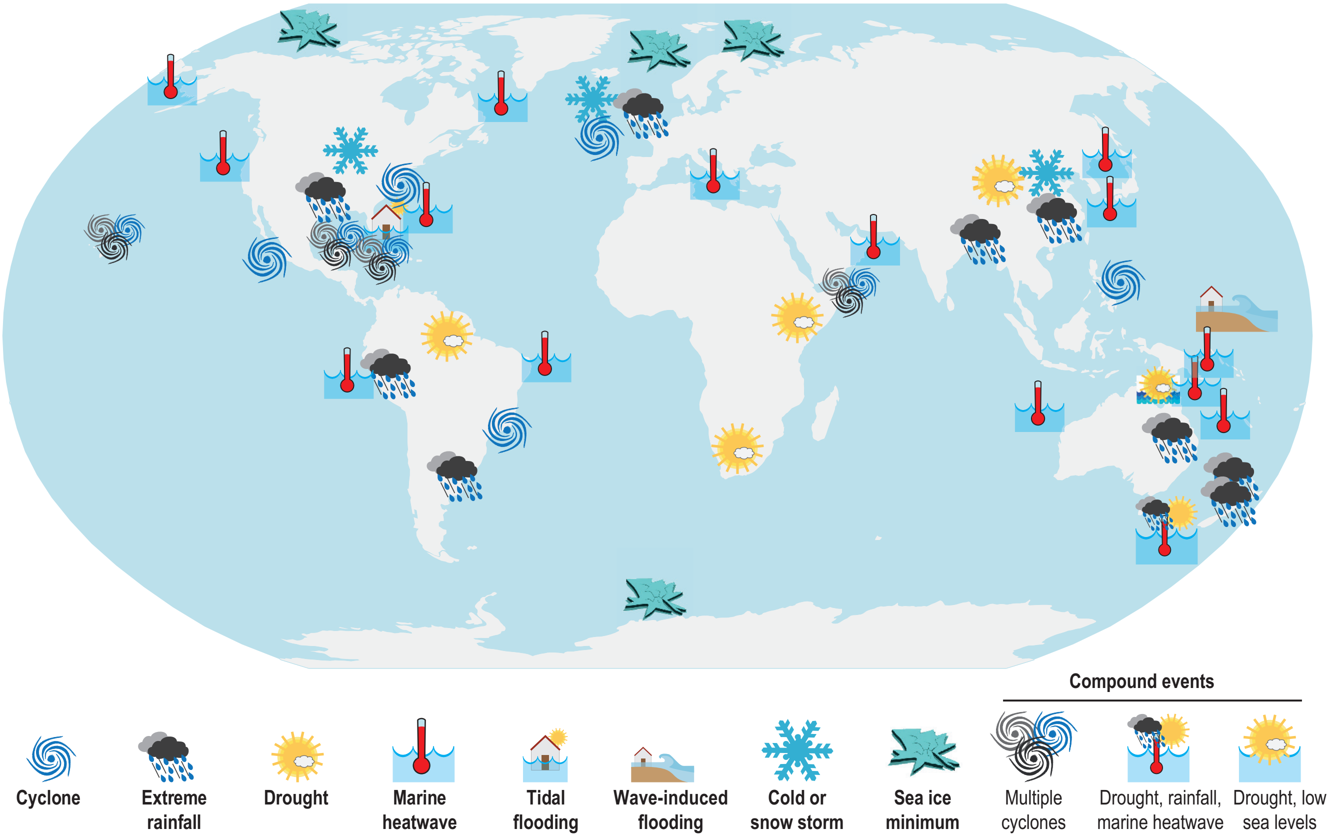

Figure 6.2 | Locations where extreme events with an identified link to ocean changes have been discussed in Table 6.2.

Table 6.2 | A selection of extreme events with links to oceans and cryosphere. In many of these studies the method of event attribution has been used to estimate the role of climate change using either a probabilistic approach (using ensembles of climate models to assess how much more likely the event has become with anthropogenic climate change compared to a world without) or a storyline approach which examines the components of the climate system that contribute to the events and how changes in the climate system affect them (Shepherd, 2016).

Year/type of hazard	Region	Severe hazard	Attribution to anthropogenic climate change	Impact, costs
1998	Western equatorial Pacific, Great Barrier Reef, Australia	Extreme sea surface temperatures (SSTs)	Unknown if global warming has increased the probability	Coral bleaching
2003	Mediterranean Sea	June to August with sea water temperatures 1°C–3°C above climatological mean (Olita et al. 2007; Garrabou et al. 2009; Galli et al. 2017)	Increase in air temperature and a reduction of wind stress and air-sea exchanges (Olita et al. 2007). Unknown if global warming has increased the probability	Mass mortality of macro-invertebrate species; amplified heatwave over central Europe in 2003
2004	South Atlantic	First hurricane in the South Atlantic since 1970	Increasing trend to positive Southern Annular Mode (SAM) could favour the synoptic conditions for such events in the future (Pezza and Simmonds, 2005)	Three deaths, 425 million USD damage (McTaggart-Cowan et al. 2006)
2005	North Atlantic	Record number of tropical storms, hurricanes and Category 5 hurricanes since 1970	Trend in SST due to global warming contributed to half of the total SST anomaly. Atlantic Multi-decadal Variability (AMV) and the after-effects of the 2004–2005 El Niño also played a role (Trenberth and Shea, 2006)	Costliest US natural disaster; 1,836 deaths and 30 billion USD in direct economic costs in Louisiana due to Hurricane Katrina (Link, 2010)
2007	Arabian Sea	Strongest tropical cyclone (TC) (Gonu) attaining sustained winds of 270 kph and gustiness of 315 kph	No attribution study done, although it was noted that this Category 5 TC had followed an unusual path (Dibajnia et al. 2010)	Caused around 4 billion USD in damages (Fritz et al. 2010; Coles et al. 2015)

Year/type of hazard	Region	Severe hazard	Attribution to anthropogenic climate change	Impact, costs
2008	Western Pacific Islands	North Pacific generated wave-swell event	Event shown to have been made more extreme compared to other historical events due to La Niña and SLR (Hoeke et al. 2013)	Wave-induced inundation in islands of six Pacific nations (Kiribati, Marshall Islands, Micronesia, Nauru, Papua New Guinea, Solomon Islands), salt water flooding of food and water supplies in Kosrae, Micronesia, 1,408 houses damaged and 63,000 people affected across eight provinces in Papua New Guinea (Hoeke et al. 2013)
2010	Western equatorial Pacific, Great Barrier Reef, Australia	Extreme SST	Unknown if global warming increased the probability	Coral bleaching
2010	Southern Amazon	Widespread drought in the Amazon led to lowest river levels of major Amazon tributaries on record (Marengo et al. 2011)	Model-based attribution indicates human influences and SST natural variability increased probabilities of the 2010 severe drought in the South Amazon region whereas aerosol emissions had little effect (Shiogama et al. 2013)	Relative to the long-term mean, the 2010 drought resulted in a reduction in biomass carbon uptake of 1.1 Pg C, compared to 1.6 Pg C for the 2005 event which was driven by an increase in biomass mortality (Feldpausch et al. 2016)
2010–2011	Eastern Australia	Wettest spring since 1900 (Leonard et al. 2014)	Based on La Niña SSTs during satellite era, La Niña alone is insufficient to explain total rainfall. 25% of rainfall was attributed to SST trend in region (Evans and Boyer-Souchet, 2012)	Brisbane river catchment flooding in January 2011, costing 23 lives and an estimated 2.55 billion USD (van den Honert and McAneney, 2011)
2010–2011	UK	Severely cold winter (coldest December since 1910 and second coldest since 1659)	Model results indicate that human influence reduced the odds by at least 20% and possibly by as much as 4 times with a best estimate that the odds have been halved (Christidis and Stott, 2012)	Many adverse consequences of the extreme temperatures, including closed schools and airports (Christidis and Stott, 2012)
2011	Western North Pacific	Tropical Storm Washi (also known as TS Sendong) was world's deadliest storm in 2011	No attribution done; disaster was the outcome of interplay of climatic, environmental and social factors (Espinueva et al. 2012)	Fatalities: >1,250; injured: 2,002; missing: 1,049 (Rasquinho et al. 2013) Socioeconomic costs: 63.3 million USD (Espinueva et al. 2012)
2011	Western Australia	Most extreme warming event in the region in the last 140 years during which sea temperature anomalies of 2°C–4°C persisted for more than 10 weeks along >2,000 km of coastline from Ningaloo (22°S) to Cape Leeuwin (34°S); up to 5°C warmer SSTs than normal (Feng et al. 2013; Pearce and Feng, 2013; Benthuysen et al. 2014; Caputi et al. 2016; Perkins-Kirkpatrick et al. 2016)	Warming of poleward-flowing Leeuwin Current in Austral summer forced by oceanic and atmospheric teleconnections associated with the 2010–2011 La Niña (Feng et al. 2013). Conditions increased since 1970's by negative Interdecadal Pacific Oscillation (IPO) and anthropogenic global warming (Feng et al. 2015). Shift of temperate marine ecosystem was climate-driven	Widespread coral bleaching and fish kills. Biodiversity patterns of temperate seaweeds, sessile invertebrates and demersal fish were altered leading to reduced abundance of habitat-forming seaweeds (Wernberg et al. 2013)
2011	Golden Bay, New Zealand	In December, Extreme two day total rainfall was experienced (one in 500-year event)	Model based attribution indicated total moisture available for precipitation in Golden Bay, New Zealand was 1–5% higher due to anthropogenic emissions (Dean et al. 2013)	In town of Takaka, 453 mm was recorded in 24 hours and 674 mm in 48 hours (Dean et al. 2013)
2012	Arctic	Arctic sea ice minimum	Model-based attribution indicated the exceptional 2012 sea ice loss was due to sea ice memory and positive feedback of warm atmospheric conditions, both contributing approximately equally (Guemas et al. 2013) and *extremely unlikely* to have occurred due to internal climate variability alone based on observations and model-based attribution (Zhang and Knutson, 2013)	Up to 60% higher contribution of sea ice algae in the central Arctic (Fernández-Méndez et al. 2015; see also chapter 3.2.3)
2012	US East coast	Hurricane Sandy	Relative SLR shown to have increased probabilities of exceeding peak impact elevations since the mid-20th century (Sweet et al. 2013; Lackmann, 2015)	Repair and mitigation expenditures funded at 60.2 billion USD. Losses of fishing vessels estimated at 52 million USD (Sainsbury et al. 2018)
2012	Northwest Atlantic	First half of 2012, record-breaking SSTs (1°C–3°C above normal) from the Gulf of Maine to Cape Hatteras (Mills et al. 2013; Chen et al. 2014; Pershing et al. 2015; Zhou et al. 2015)	Local warming from the atmosphere due to anomalous atmospheric jet stream position (Chen et al. 2014). Unknown if global warming increased the probability	Northward movement of warm water species and local migrations of lobsters earlier in the season (Mills et al. 2013; Pershing et al. 2015)

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Year/type of hazard	Region	Severe hazard	Attribution to anthropogenic climate change	Impact, costs
2013	UK	Extreme winter rainfall	Some evidence for a human-induced increase in extreme winter rainfall in the UK for events with time scales of 10 days (Christidis and Stott, 2015)	Tidal surges, widespread floodplain inundation, and pronounced river flows leading to damages in transport infrastructure, business and residential properties and a cost of 560 million GBP in recovery schemes (Department for Communities and Local Government, 2014; Huntingford et al. 2014). Unprecedented deaths of over 4,400 Puffins found on UK and Scottish coasts linked to cold and strong winds during this event (Harris and Elkins, 2013)
2013	Western North Pacific	Strongest and fastest Super Typhoon Haiyan (Category 5) in the region	Occurred in a season with remarkably warm SSTs, (David et al. 2013; Takagi and Esteban, 2016). Ocean heat content and sea levels had increased since 1998 due to the negative Pacific Decadal Oscillation (PDO) phase but impacts were worsened by thermodynamic effects on SSTs, SLR and storm surges due to climate change (Trenberth et al. 2015)	Deadliest and most expensive natural disaster in the Philippines (Fatalities: 6,245; Injured: 28,626; Missing: 1,039). Damage to mangroves was still apparent 18 months after the storm (Sainsbury et al. 2018)
2013–2015	Northeast Pacific Ocean	Largest heatwave ever recorded (often called 'The Blob'; Bond et al. 2015), with maximum SST anomalies of 6°C off Southern California (Jacox et al. 2016; Gentemann et al. 2017; Rudnick et al. 2017) and subsurface warm anomalies in the deep British Columbia Fjord that persisted through the beginning of 2018 (Jackson and Wood, 2018)	Emerged in 2013 in response to teleconnections between North Pacific and the weak El Niño that drove strong positive sea level pressure anomalies across the northeast Pacific inducing smaller heat loss (Bond et al. 2015; Di Lorenzo and Mantua, 2016). Global warming increased the probability of occurrence for regional parts of the MHW (Weller et al. 2015; Jacox et al. 2018; Newman et al. 2018)	Major impacts on entire marine food web. Caused a major outbreak of a toxic algal bloom along the US West Coast leading to impacts on fisheries (McCabe et al. 2016) . Increased mortality of sea birds (Jones et al. 2018). Contributed to drought conditions across the US West Coast
2014	Hawaiian hurricane season	Extremely active hurricane season in the eastern and central Pacific Ocean, particularly around Hawaii	Anthropogenic forcing could have contributed to the unusually large number of hurricanes in Hawaii in 2015, in combination with the moderately favourable El Niño event conditions (Murakami et al. 2015)	Acute disturbance of coral along Wai'ōpae coastline (southeastern tip of Hawai'i Island) due to passages of Hurricanes Iselle, Julio and Ana that caused high waves, increased runoff and elevated SSTs associated with the 2014–2015 El Niño (Burns et al. 2016)
2014	Arabian Sea	Cyclone Nilofar was the first severe TC to be recorded in the Arabian Sea in post-monsoon cyclone season (Murakami et al. 2017)	Anthropogenic global warming has been shown to have increased the probability of post-monsoon TCs over the Arabian Sea (Murakami et al. 2017)	Cyclone did not make landfall but produced heavy rainfall on western Indian coasts (Bhutto et al. 2017)
2014	Northland New Zealand	Extreme five day rainfall in Northland	Extreme five day rainfall over Northland, New Zealand was influenced by human-induced climate change (Rosier et al. 2015)	18.8 million NZD in insurance claims (Rosier et al. 2015)
2014–2017	Western equatorial Pacific, Great Barrier Reef, Australia	Extreme SSTs	Global warming increased probability of occurrence for regional parts of the MHW (Weller et al. 2015; Oliver et al. 2018b)	Anthropogenic greenhouse gas (GHG) emission increased the risk of coral bleaching through anomalously high SSTs and accumulation of heat stress (Lewis and Mallela, 2018)
2015	North America	Anomalously low temperatures with intense snowstorms	Reduced Arctic sea ice and anomalous SSTs may have contributed to establishing and sustaining the anomalous meander of the jet stream, and could enhance the probability of such extreme cold spells over North America (Bellprat et al. 2016)	Several intense snowstorms resulting in power outages and large economic losses (Munich RE, 2016)
2015	Arctic	Record low Northern Hemisphere (NH) sea ice extent in March 2015	Record low in NH sea ice maximum could not have been reached without human-induced change in climate, with the surface atmospheric conditions, on average, contributing 54% to the change (Fuckar et al. 2016)	March NH sea ice content reached the lowest winter maximum in 2015. Emerging evidence of increased snow fall over regions outside the Arctic (see 3.4.1.1) due to sea ice reduction as well as changes in the timing, duration and intensity of primary production, which affect secondary production (3.2.3.1)
2015	Florida	Sixth largest flood in Virginia Key, Florida since 1994, with the fifth highest in response to hurricanes	The probability of a 0.57 m flood has increased by 500% (Sweet et al. 2016)	Flooding in several Miami-region communities with 0.57 m of ocean water on a sunny day

Year/type of hazard	Region	Severe hazard	Attribution to anthropogenic climate change	Impact, costs
2015–2016	Ethiopia and Southern Africa	One of the worst droughts in 50 years, also intensified flash droughts characterised by severe heatwaves	Anthropogenic warming contributed substantially to the very warm 2015–2016 El Niño SSTs, land local air temperatures thereby reducing Northern Ethiopia and Southern Africa rainfall and runoff (Funk et al. 2018; Yuan et al. 2018)	A 9 million tonne cereal deficit resulted in more than 28 million people in need of humanitarian aid (Funk et al. 2018)
2015	Eastern North Pacific	TC Patricia, the most intense and rapidly intensifying storm in the Western Hemisphere (estimated mean sea level (MSL) pressure of 872 hPa (Rogers et al. 2017), intensified rapidly into a Category 5 TC (Diamond and Schreck, 2016)	A near-record El Niño combined with a positive Pacific Meridional Mode provided extreme record SSTs and low vertical wind shear that fuelled the 2015 eastern North Pacific hurricane season to near-record levels (Collins et al. 2016)	Approximately 9,000 homes and agricultural croplands, including banana crops, were damaged by wind and rain from Patricia that made landfall near Jalisco, Mexico (Diamond and Schreck, 2016)
2015	Arabian Sea, Somalia and Yemen	Cyclones Chapala and Megh occurred within a week of each other and both tracked westward across Socotra Island and Yemen. Rainfall from Chapala was seven times the annual average	Anthropogenic global warming has been shown to have increased the probability of post-monsoon TCs over the Arabian Sea (Murakami et al. 2017)	Death toll in Yemen from Chapala and Megh was 8 and 20 respectively. Thousands of houses and businesses damaged or destroyed by both cyclones and fishing disrupted. The coastal town of Al Mukalla experienced a 10 m storm surge that destroyed the seafront (Kruk, 2016). Flooding in Somalia led to thousands of livestock killed and damage to infrastructure (IFRC, 2016)
2015–2016	Northern Australia (Gulf of Carpentaria)	High temperatures, low rainfall, extended drought period and low sea levels	Attributed to anomalously high temperatures and low rainfall and low sea levels associated with El Niño (Duke et al. 2017)	1,000 km of mangrove tidal wetland dieback (>74,000 ha). with potential flow-on consequences to Gulf of Carpentaria fishing industry worth 30 million AUS yr^{-1} due to loss of recruitment habitat
2015–2016	Tasman Sea	MHW lasted for 251 days with maximum SSTs of 2.9°C above the 1982–2005 average (Oliver et al. 2017)	Enhanced southward transport in the East Australian current driven by increased wind stress (Oliver et al. 2017). The intensity and duration of the MHW were unprecedented and both had a clear human signature (Oliver et al. 2017)	Disease outbreaks in farmed shellfish, mortality in wild shellfish and species found further south than previously recorded. Drought followed by severe rainfall caused severe bushfires and flooding in northeast Tasmania (see Box 6.1)
2016	Arctic	Record high air temperatures and record low sea ice were observed in the Arctic winter/spring of 2016 (Petty et al. 2017)	Would not have been possible without anthropogenic forcing (Kam et al. 2018), however the relative role of preconditioning, seasonal atmospheric/ocean forcing and storm activity in determining the evolution of the Arctic sea ice cover is still highly uncertain (Petty et al. 2018)	Impacts on Arctic ecosystems (e.g., Post et al. 2013; Meier et al. 2014), potential changes to mid-latitude weather (e.g., Cohen et al. 2014; Francis and Skific, 2015; Screen et al. 2015) and human activities in the Arctic
2016	Bering Sea/Gulf of Alaska	Record-setting warming with peak SSTs of 6°C above the 1981–2010 climatology (Walsh et al. 2017; Walsh et al. 2018)	Nearly fully attributed to human-induced climate change (Oliver et al. 2018b; Walsh et al. 2018)	Impacts on marine ecosystems in Alaska, included favouring some phytoplankton species, but resulted in one of the largest harmful algal blooms on record which reached the Alaska coast in 2015 (Peterson et al. 2017), uncommon paralytic shellfish poisoning events in Kachemak Bay and oyster farm closures in 2015 and 2016, dramatic mortality events in seabird species such as common murres in 2015–2016 (Walsh et al. 2018)
2016	East China Sea	MHW	Warming predominantly attributable to combined effects of oceanic advection (-0.18°C, 24%) and net heat flux (-0.44°C, 58%; Tan and Cai, 2018)	Impacts on marine organisms (Kim and Han, 2017)
2016	Eastern China	Super cold surge	This cold surge would have been stronger if there was no anthropogenic warming (Qian et al. 2018; Sun and Miao, 2018)	Extreme weather brought by the cold surge caused significant impacts on >1 billion people in China in terms of transportation and electricity transmission systems, agriculture and human health (Qian et al. 2018)
2016	Antarctic	Antarctic sea ice extent decreased at a record rate 46% faster than the mean rate and 18% faster than any spring rate in the satellite era producing a record minimum for the satellite period (1979–2016) (Turner et al. 2017)	Largely attributable to thermodynamic surface forcing (53%), while wind stress and the sea ice and oceanic conditions from the previous summer (January 2016) explain the remaining 34% and 13%, respectively (Kusahara et al. 2018) linked with a shift to positive phase of PDO and negative SAM in late 2016 (Meehl et al. 2019; see also 3.2.1.1)	Potential impacts on ecosystems and fisheries are poorly known (chapter 3.6)

Year/type of hazard	Region	Severe hazard	Attribution to anthropogenic climate change	Impact, costs
2017	Yellow Sea/ East China Sea	SSTs 2°C–7°C higher than normal (Kim and Han, 2017; Tan and Cai, 2018)	Unknown if global warming has increased the probability	Impacts on marine organisms
2017	Western North Atlantic	Hurricanes Harvey, Irma and Maria	Rainfall intensity in Harvey attributed to climate change and winds for Irma and Maria attributed to climate change. (Emanuel, 2017; Risser and Wehner, 2017; van Oldenborgh et al. 2017; see Box 6.1)	Extensive impacts (see Box 6.1)
2017	Europe	Storm Ophelia	In agreement with projections of increase of cyclones of tropical origin hitting European coasts (Haarsma et al. 2013)	Largest ever recorded hurricane in East Atlantic; extreme winds and coastal erosion in Ireland
2017	Persian Gulf	Severe warming in the Gulf with reef bottom temperatures resulting in 5.5°C-weeks of thermal stress as degree heating weeks (Burt et al. 2019)	Mortality of corals shown to have been caused by increases in sea-bottom temperatures (Burt et al. 2019)	94.3% of corals bleached in the Gulf
2017	East Africa	Drought (across Tanzania, Ethiopia, Kenya and Somalia)	Extremely warm 'Western V' (stretching poleward and eastward from a point near the Maritime Continent) SST doubled the probability of drought (Funk et al. 2018)	Contributed to extreme food insecurity (Funk et al. 2018) approaching near-famine conditions (FEWS NET and FSNAU, 2017; WFP et al. 2017)
2017	Peru	Extremely wet rainy season	Human influence is estimated to make such events at least 1.5 times more likely (Christidis et al. 2018a)	Widespread flooding and landslides 1.7 million people, a death toll of 177 and an estimated damages of 3.1 billion USD (Christidis et al. 2018a)
2017	Bangladesh	Pre-monsoon extreme six day rainfall event	The likelihood of this 2017 pre-monsoon extreme rainfall is nearly doubled by anthropogenic climate change; although this contribution is sensitive to the climatological period used (Rimi et al. 2018)	Triggered flash floods affecting 850,000 house-holds and 220,000 hectares of harvestable crops leading to a 30% rice price hike (FAO, 2017)
2017	Uruguay, South America	April-May heavy precipitation	The risk of the extreme rainfall in the Uruguay River increased two-fold by anthropogenic climate change	Triggered wide-spread overbank flooding along the Uruguay River causing economic loss of 102 million USD (FAMURS, 2017) and displace-ment of 3,500 people (de Abreu et al. 2019)
2017	Northeast China	Persistent summer-spring hot and dry extremes	Risk of persistent spring-summer hot and dry extremes is increased by 5–55% and 37–113%, respectively, by anthropogenic climate change (Wang et al. 2018)	Affected more than 7.4 million km² of crops and herbage and direct economic loss of about 10 billion USD (Zhang et al. 2017c)
2017	Coastal Peru	Strong shallow ocean warming of up to 10°C off the northern coast of Peru	Unknown if global warming increased the probability	Caused heavy rainfall and flooding (ENFEN, 2017; Garreaud, 2018). Affected anchovies (decreased fat content and early spawning as a reproductive strategy; IMPARPE, 2017)
2017	Southwestern Atlantic	SSTs were 1.7°C higher than previous maximum from February to March 2017 between 32°S–38°S (Manta et al. 2018)	High air temperature and low wind speed led to MHW. Unknown if global warming increased the probability	Fish species mass mortalities

6.3 Changes in Tracks, Intensity, and Frequency of Tropical and Extratropical Cyclones and Associated Sea Surface Dynamics

This section addresses new literature on TCs and ETCs and their effects on the ocean in the context of understanding how the changing nature of extreme events can cause compound hazards, risk and cascading impacts (discussed in Section 6.8). These topics are also discussed in Chapter 4 in the context of changes to ESLs (see Section 4.2.3.4).

6.3.1 Changes in Storms and Associated Sea Surface Dynamics

6.3.1.1 Tropical Cyclones

IPCC AR5 concluded that there was *low confidence* in any long-term increases in TC activity globally and in attribution of global changes to any particular cause (Bindoff et al. 2013; Hartmann et al. 2013). Based on process understanding and agreement in 21st century projections, it is *likely* that the global TC frequency will either decrease or remain essentially unchanged, while global mean

TC maximum wind speed and precipitation rates will *likely* increase although there is *low confidence* in region-specific projections of frequency and intensity (Christensen et al. 2013). The AR5 concluded that circulation features have moved poleward since the 1970s, associated with a widening of the tropical belt, a poleward shift of storm tracks and jet streams, and contractions of the northern polar vortex and the Southern Ocean westerly wind belts. However it is noted that natural modes of variability on interannual to decadal time scales prevent the detection of a clear climate change signal (Hartmann et al. 2013).

Since the AR5 and Knutson et al. (2010), palaeoclimatic surveys of coastal overwash sediments and stalagmites have provided further evidence of historical TC variability over the past several millennia. Patterns of storm activity across TC basins show variations through time that appear to be correlated with El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), and changes in atmospheric dynamics related to changes in precession of the sun (Toomey et al. 2013; Denommee et al. 2014; Denniston et al. 2015).

Further studies have investigated the dynamics of TCs. A modelling study investigated a series of low-frequency increases and decreases in TC activity over the North Atlantic over the 20th century (Dunstone et al. 2013). These variations, culminating in a recent rise in activity, are thought to be due in part to atmospheric aerosol forcing variations (aerosol forcing), which exerts a cooling effect (Booth et al. 2012; Dunstone et al. 2013). However, the relative importance of internal variability vs. radiative forcing for multidecadal variability in the Atlantic basin, including TC variability, remains uncertain (Weinkle et al. 2012; Zhang et al. 2013; Vecchi et al. 2017; Yan et al., 2017). Although the aerosol cooling effect has largely cancelled the increases in potential intensity over the observational period, according to Coupled Model Intercomparison Project Phase 5 (CMIP5) model historical runs, further anthropogenic warming in the future is expected to dominate the aerosol cooling effect leading to increasing TC intensities (Sobel et al. 2016).

TCs amplify wave heights along the tracks of rapidly moving cyclones (e.g., Moon et al. 2015a) and can therefore increase mixing to the surface of cooler subsurface water. Several studies found that TCs reduce the projected thermal stratification of the upper ocean in CMIP5 models under global warming, thereby slightly offsetting the simulated TC-intensity increases under climate warming conditions (Emanuel, 2015; Huang et al. 2015b; Tuleya et al. 2016). On the other hand, freshening of the upper ocean by TC rainfall enhances density stratification by reducing near-surface salinity and this reduces the ability of TC's to cool the upper ocean, thereby having an influence opposite to the thermal stratification effect (Balaguru et al. 2015). In the late 21st century, increased salinity stratification was found to offset about 50% of the suppressive effects that TC mixing has on temperature stratification (Balaguru et al. 2015). Coupled ocean-atmosphere models still robustly project an increase of TC intensity with climate warming, and particularly for new TC-permitting coupled climate model simulations that compute internally consistent estimates of thermal stratification change (e.g., Kim et al. 2014a; Bhatia et al. 2018). Higher TC intensities in turn may further aggravate the impacts of SLR on TC-related coastal inundation extremes (Timmermans et al. 2017).

Kossin et al. (2014) identified a poleward expansion of the latitudes of maximum TC intensity in recent decades, which has been linked to an anthropogenically-forced tropical expansion (Sharmila and Walsh, 2018) and a continued poleward shift of cyclones projected over the western North Pacific in a warmer climate (Kossin et al. 2016). A 10% slowdown in translation speed of TCs over the 1949–2016 period has been linked to the weakening of the tropical summertime circulation associated with tropical expansion and a more pronounced slowdown in the range 16–22% was found over land areas affected by TCs in the western North Pacific, North Atlantic and Australian regions (Kossin, 2018). Slow-moving TCs together with higher moisture carrying capacity can cause significantly greater flood hazards (Emanuel, 2017; Risser and Wehner, 2017; van Oldenborgh et al. 2017; see also Table 6.2 and Box 6.1).

Trends in TCs over decades to a century or more have been investigated in several new studies. Key findings include: i) decreasing frequency of severe TCs that make landfall in eastern Australia since the late 1800s (Callaghan and Power, 2011); ii) increase in frequency of moderately large US storm surge events since 1923 (Grinsted et al. 2012); iii) recent increase of extremely severe cyclonic storms over the Arabian Sea in the post-monsoon season (Murakami et al. 2017); iv) intense TCs that make landfall in East and Southeast Asia in recent decades (Mei and Xie, 2016; Li et al. 2017); and v) an increase in annual global proportion of hurricanes reaching Category 4 or 5 intensity in recent decades (Holland and Bruyère, 2014).

Rapid intensification of tropical cyclones (RITCs) poses forecast challenges and increased risks for coastal communities (Emanuel, 2017). Warming of the upper ocean in the central and eastern tropical Atlantic associated with the positive phase of the Atlantic Multidecadal Oscillation (AMO) (Balaguru et al. 2018) and in the western North Pacific in recent decades due to a La Niña-like pattern (Zhao et al. 2018) has favoured RITCs in these regions. One new modelling study suggests there has been a detectable increase in RITC occurrence in the Atlantic basin in recent decades, with a positive contribution from anthropogenic forcing (Bhatia et al. 2019). Nonetheless, the background conditions that favour RITC's across the Atlantic basin as a whole tend to be associated with less favourable conditions for TC occurrence along the US east coast (Kossin, 2017).

New studies have used event attribution to explore attribution of certain individual TC events or anomalous seasonal cyclone activity events to anthropogenic forcing (Lackmann, 2015; Murakami et al. 2015; Takayabu et al. 2015; Zhang et al. 2016; Emanuel, 2017; see also Table 6.2 and Box 6.1). Risser and Wehner (2017) and van Oldenborgh et al. (2017) concluded that for the Hurricane Harvey event, there is a detectable human influence on extreme precipitation in the Houston area, although their detection analysis is for extreme precipitation in general and not specifically for TC-related precipitation.

There have been more TC dynamical or statistical/dynamical downscaling studies and higher resolution General Circulation Model (GCM) experiments (e.g., Emanuel, 2013; Manganello et al. 2014; Knutson et al. 2015; Murakami et al. 2015; Roberts et al. 2015; Wehner et al. 2015; Yamada et al. 2017). The findings of these studies generally support the AR5 projections of a general increase in intensity of the most intense TCs and a decline in TC frequency overall. However, the projected increase in global TC frequency by Emanuel (2013) and Bhatia et al. (2018) differed from most other TC frequency projections and previous assessments. For studies into future track changes of TCs under climate warming scenarios (Li et al. 2010; Kim and Cai, 2014; Manganello et al. 2014; Knutson et al. 2015; Murakami et al. 2015; Roberts et al. 2015; Wehner et al. 2015; Nakamura et al. 2017; Park et al. 2017; Sugi et al. 2017; Yamada et al. 2017; Yoshida et al. 2017; Zhang et al. 2017a), it is difficult to identify a robust consensus of projected change in TC tracks, although several of the studies found either poleward or eastward expansion of TC occurrence over the North Pacific region resulting in greater storm occurrence in the central North Pacific. There have been new studies on storm size (Kim et al. 2014a; Knutson et al. 2015; Yamada et al. 2017) under climate warming scenarios. These project TC size changes of up to ±10% between basins and studies and provide preliminary findings on this issue that future studies will continue to investigate. Several studies of TC storm surge (e.g., Lin et al. 2012; Garner et al. 2017) suggest that SLR will dominate the increased height of storm surge due to TCs under climate change.

Taking the above into account, the following is a summary assessment of TC detection and attribution. The observed poleward migration of the latitude of maximum TC intensity in the western North Pacific appears to be unusual compared to expected natural variability and therefore there is *low to medium confidence* that this change represents a detectable climate change, though with only *low confidence* that the observed shift has a discernible positive contribution from anthropogenic forcing. Anthropogenic forcing is believed to be producing some poleward expansion of the tropical circulation with climate warming. Additional studies of observed long-term TC changes such as: an increase in annual global proportion of Category 4 or 5 TCs in recent decades, severe TCs occurring in the Arabian Sea, TCs making landfall in East and Southeast Asia, the increasing frequency of moderately large US storm surge events since 1923 and the decreasing frequency of severe TCs that make landfall in eastern Australia since the late 1800s, may each represent emerging anthropogenic signals, but still with *low confidence* (*limited evidence*). The lack of confident climate change detection for most TC metrics continues to limit confidence in both future projections and in the attribution of past changes and TC events, since TC event attribution in most published studies is generally being inferred without support from a confident climate change detection of a long-term trend in TC activity.

TCs projections for the late 21st century are summarised as follows: 1) there is *medium confidence* that the proportion of TCs that reach Category 4–5 levels will increase, that the average intensity of TCs will increase (by roughly 1–10%, assuming a 2°C global temperature rise), and that average TCs precipitation rates (for a given storm) will increase by at least 7% per degree Celsius SST warming, owing to higher atmospheric water vapour content, 2) there is *low confidence* (*low agreement*, *medium evidence*) in how global TC frequency will change, although most modelling studies project some decrease in global TC frequency and 3) SLR will lead to higher storm surge levels for the TCs that do occur, assuming all other factors are unchanged (*very high confidence*).

6.3.1.2 Extratropical Cyclones and Blocking

ETCs form in the mid-latitudes of the North Atlantic, North Pacific and Southern Oceans, and the Mediterranean Sea. The storm track regions are characterised by large surface equator-to-pole temperature gradients and baroclinic instability, and jet streams influence the direction and speed of movement of ETCs in this region. The thermodynamic response of the atmosphere to CO_2 tends to have opposing influences on storm tracks; surface shortwave cloud radiative changes increase the equator-to-pole temperature gradient whereas longwave cloud radiative changes reduce it (Shaw et al. 2016). AR5 concluded that the global number of ETCs is not expected to decrease by more than a few percent due to anthropogenic change. The Southern Hemisphere (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).

A 'blocking' event is an extratropical weather system in which the anticyclone (region of high pressure) becomes quasi-stationary and interrupts the usual westerly flow and/or storm tracks for up to a week or more (Woollings et al. 2018). Recent attention has focused on whether Arctic warming is linked to increased blocking and mid-latitude weather extremes (Barnes and Screen, 2015; Francis and Skific, 2015; Francis and Vavrus, 2015; Kretschmer et al. 2016), such as drought in California due to sea ice changes that cause a reorganisation of tropical convection (Cvijanovic et al. 2017), cold and snowy winters over Europe and North America (Liu et al. 2012; Cohen et al. 2018), extreme summer weather (Tang et al. 2013; Coumou et al. 2014) and Balkan flooding (Stadtherr et al. 2016). Studies suggest how blocking may influence arctic sea ice extent (Gong and Luo, 2017) and various pathways whereby Arctic warming could influence extreme weather (Barnes and Screen, 2015) such as reducing the equator to pole temperature gradient, slowing the jet stream thereby increasing its meandering behaviour (Röthlisberger et al. 2016; Mann et al. 2017) or causing it to split (Coumou et al. 2014), changing local dynamics in the vicinity of the sea ice edge (Screen and Simmonds, 2013) or weakening the stratospheric polar vortex (Cohen et al. 2014). However, sensitivity to choice of methodology (Screen and Simmonds, 2013) and large internal atmospheric variability masks the detection of such links in past records, and climate change can lead to opposing effects on the mid-latitude jet stream response leading to large uncertainty in future changes (Barnes and Polvani, 2015; Barnes and Screen, 2015).

New studies of future storm track behaviour in the NH, include Harvey et al. (2014) who find that the future changes to upper and lower tropospheric equator-to-pole temperature differences by the end of the century in a CMIP5 multi-model RCP8.5 ensemble are not well correlated and the lower temperature gradient dominates the summer storm track response whereas both upper and lower temperature gradients play a role in winter. In the northern North Atlantic storm track region, projected changes are found to be more strongly associated with changes in the lower rather than upper tropospheric equator-to-pole temperature difference (Harvey et al. 2015). In the SH, Harvey et al. (2014) find equator-to-pole temperature differences in the upper and lower troposphere in the future climate across a multi-model ensemble are well correlated with a general strengthening of the storm track. The total number of ETCs in a CMIP5 GCM multi-model ensemble decreased in the future climate, whereas the number of strong ETCs increased in most models and in the ensemble mean (Grieger et al. 2014). This was associated with a general poleward shift related to both tropical upper tropospheric warming and shifting meridional SST gradients in the Southern Ocean. The poleward movement of baroclinic instability and associated storm formation over the observational period due to external radiative forcing, is projected to continue, with associated declining rainfall trends in the mid-latitudes and positive trends further polewards (Frederiksen et al. 2017).

A number of new studies have found links between Arctic amplification, blocking events and various types of weather extremes in NH mid-latitudes in recent decades. However, the sensitivity of results to analysis technique and the generally short record with respect to internal variability means that at this stage there is *low confidence* in these connections. Consistent with the AR5, projected changes to NH storm tracks exhibit large differences between responses, causal mechanisms and ocean basins and so there remains *low confidence* in future changes in blocking and storm tracks in the NH. The storm track projections for the SH remain consistent with previous studies in indicating an observed poleward contraction and a continued strengthening and southward contraction of storm tracks in the future (*medium confidence*).

6.3.1.3 Waves and Extreme Sea Levels

AR5 also concluded that there is *medium confidence* that mean significant wave height has increased in the North Atlantic north of 45°N based on ship observations and reanalysis-forced wave model hindcasts. ESL events have increased since 1970, mainly due to a rise in mean sea levels (MSLs) over this period (Rhein et al. 2013). There is *medium confidence* that mid-latitude jets will move 1–2 degrees further poleward by the end of the 21st century under RCP8.5 in both hemispheres with weaker shifts in the NH. In the SH during austral summer, the poleward movement of the mid-latitude westerlies under climate change is projected to be partially offset by stratospheric ozone recovery. There is *low confidence* in projections of NH storm tracks particularly in the North Atlantic. Tropical expansion is *likely* to continue causing wider tropical regions and poleward movement of the subtropical dry zones (Collins et al. 2013). In the SH, it is *likely* that enhanced wind speeds will cause an increase in annual mean significant wave heights. Wave swells generated in the Southern Ocean may also affect wave heights, periods and directions in adjacent ocean basins. The projected reduction in sea ice extent in the Arctic Ocean (Holland et al. 2006) will increase wave heights and wave season length (Church et al. 2013).

Since AR5, new studies have shown observed changes in wave climate. Satellite observations from 1985–2018, showed small increases in significant wave height (+0.3 cm/year) and larger increases in extreme wave heights (90th percentiles), especially in the Southern (+1 cm/year) and North Atlantic (+0.8 cm/year) Oceans (Young and Ribal, 2019) as well as positive trends in wave height in the Arctic over 1992–2014 due to sea ice loss (Stopa et al. 2016; Thomson et al. 2016). Based on a wave reanalysis and satellite observations, Reguero et al. (2019) found that the global wave power, which represents the transport of the energy transferred from the wind into the sea surface motion, therefore including wave height, period and direction, has increased globally at a rate of 0.41% yr^{-1} between 1948 and 2008, with large variations across oceans. Long-term correlations are found between the increase in wave power and SSTs, particularly between the tropical Atlantic temperatures and the wave power in high southern latitudes, the most energetic region globally.

The results of several new global wave climate projection studies are consistent with those presented in IPCC AR5. Mentaschi et al. (2017) find up to a 30% increase in 100-year return level wave energy flux (the rate of transfer of wave energy) for the majority of coastal areas in the southern temperate zone, and a projected decrease in wave energy flux for most NH coastal areas at the end of the century in wave model simulations forced by six CMIP5 RCP8.5 simulations. The most significant long-term trends in extreme wave energy flux are explained by their relationship to modelled climate indices (Arctic Oscillation, ENSO and NAO). Wang et al. (2014b) assessed the climate change signal and uncertainty in a 20-member ensemble of wave height simulations, and found model uncertainty (inter-model variability) is significant globally, being about 10 times as large as the variability between RCP4.5 and RCP8.5 scenarios. In a study focussing on the western north Pacific wave climate, Shimura et al. (2015) associate projected regions of future change in wave climate with spatial variation of SSTs in the tropical Pacific Ocean. A review of 91 published global and regional scale wind-wave climate projection studies found a consensus on a projected increase in significant wave height over the Southern Ocean, tropical eastern Pacific (*high confidence*) and Baltic Sea (*medium confidence*), and decrease over the North Atlantic and Mediterranean Sea. They found little agreement between studies of projected changes over the Atlantic Ocean, southern Indian and eastern North Pacific Ocean and no regional agreement of projected changes to extreme wave height. It was noted that few studies focussed on wave direction change, which is important for shoreline response (Morim et al. 2018).

Significant developments have taken place since the AR5 to model storm surges and tides at the global scale. An unstructured global hydrodynamic modelling system has been developed with maximum coastal resolution of 5 km (Verlaan et al. 2015) and used to develop a global climatology of ESLs due to the combination of storm surge and tide (Muis et al. 2016). A global modelling study finds that under SLR of 0.5–10 m, changes to astronomical tidal mean high water exceed the imposed SLR by 10% or more at around 10% of coastal cities when coastlines are held fixed. When coastal recession is permitted a reduction in tidal range occurs due to changes in the period of oscillation of the basin under the changed coastline morphology (Pickering et al. 2017). A recent study on global probabilistic projections of ESLs considering MSL, tides, wind-waves and storm surges shows that under RCP4.5 and RCP8.5, the global average 100-year ESL is *very likely* to increase by 34–76 cm and 58–172 cm, respectively between 2000–2100 (Vousdoukas et al. 2018). Despite the advancements in global tide and surge modelling, using CMIP GCM multi-model ensembles to examine the effects of future weather and circulation changes on storm surges in a globally consistent way is still a challenge because of the *low confidence* in GCMs being able to represent small scale weather systems such as TCs. To date only a small number of higher resolution GCMs are able to produce credible cyclone climatologies (e.g., Murakami et al. 2012) although this will probably improve with further GCM development and increases to GCM resolution (Walsh et al. 2016).

The role of austral winter swell waves on ESL have been investigated in the Gulf of Guinea (Melet et al. 2016) and the Maldives (Wadey et al. 2017). Multivariate statistical analysis and probabilistic modelling is used to show that flood risk in the northern Gulf of Mexico is higher than determined from short observational records (Wahl et al. 2016). In Australia, changes in ESLs were modelled using four CMIP5 RCP8.5 simulations (Colberg et al. 2019). On the southern mainland coast, the southward movement of the subtropical ridge in the climate models led to small reductions (up to 0.4 m) in the modelled 20-year (5% probability of occurring in a year) storm surge. Over the Gulf of Carpentaria in the north, changes were largest and positive during austral summer in two out of the four models in response to a possible eastward shift in the northwest monsoon. Synthetic cyclone modelling was used to evaluate probabilities, interannual variability and future changes of extreme water levels from tides and TC-induced storm surge (storm tide) along the coastlines of Fiji (McInnes et al. 2014) and Samoa (McInnes et al. 2016). Higher resolution modelling for Apia, Samoa incorporating waves highlights that although SLR reduces wave setup and wind setup by 10–20%, during storm surges it increases wave energy reaching the shore by up to 200% (Hoeke et al. 2015).

In the German Bight, Arns et al. (2015) show that under SLR, increases in extreme water levels occur due to a change in phase of tidal propagation; which more than compensates for a reduction in storm surge due to deeper coastal sea levels. Vousdoukas et al. (2017) develop ESL projections for Europe that account for changes in waves and storm surge. In 2100, increases of up to 0.35 m relative to the SLR projections occur towards the end of the century under RCP8.5 along the North Sea coasts of northern Germany and Denmark and the Baltic Sea coast, whereas little to negative change is found for the southern European coasts.

In the USA, Garner et al. (2017) combine downscaled TCs, storm surge models, and probabilistic SLR projections to assess flood hazard associated with changing storm characteristics and SLR in New York City from the pre-industrial era to 2300. Increased storm intensity was found to compensate for offshore shifts in storm tracks leading to minimal change in modelled storm surge heights through 2300. However, projected SLR leads to large increases in future overall flood heights associated with TCs in New York City. Consequently, flood height return periods that were ~500y (0.2% probability of occurring in a given year) during the pre-industrial era have fallen to ~25y (4% probability of occurring annually) at present and are projected to fall to ~5y (20% probability of occurring annually) within the next three decades.

In summary, new studies on observed wave climate change from 1985–2018 showed small increases in significant wave height of +0.3 cm/year and larger increases in 90th percentile wave heights of +1 cm/year in the Southern Ocean and +0.8 cm/year in the North Atlantic ocean (*medium confidence*). Sea ice loss in the Arctic has also increased wave heights over the period 1992–2014 (*medium confidence*). Global wave power has increased over the last six decades with differences across oceans related to long-term correlations with SST (*low confidence*). Future projections indicate an increase of the mean significant wave height across the Southern Ocean and tropical eastern Pacific (*high confidence*) and Baltic Sea (*medium confidence*) and decrease over the North Atlantic and Mediterranean Sea under RCP8.5 (*high confidence*). Extreme waves are projected to increase in the Southern Ocean and decrease in the North Atlantic and Mediterranean Sea under RCP4.5 and RCP8.5 (*high confidence*). There is still limited knowledge on projected wave period and direction. For coastal ESLs, new studies at the regional to global scale have generally had a greater focus on multiple contributing factors such as waves, tides, storm surges and SLR. At the global scale, probabilistic projections of extreme sea levels considering these factors projects the global average 100-year ESL is *very likely* to increase by 34–76 cm and 58–172 cm, under RCP4.5 and RCP8.5, respectively between 2000–2100.

6.3.2 Impacts

As shown in previous assessments, increasing exposure is a major driver of increased cyclone risk (wind damages), as well as flood risk associated with cyclone rainfall and surge, besides possible changes in hazard intensities from anthropogenic climate change (Handmer et al. 2012; Arent et al. 2014). Changes in TC trajectories are potentially a major source of increased risk, as the degree of vulnerability is typically much higher in locations that were previously not exposed to the hazard (Noy, 2016). Typhoon Haiyan's move to the south of the usual trajectories of TCs in the western North Pacific basin (Yonson et al. 2018) made the evacuation more difficult as people were less willing to heed storm surge warnings they received.

Abrupt changes in impacts therefore are not only determined by changes in cyclone hazard, but also by the sensitivity or tipping points that are crossed in terms of flooding for instance, that can be driven by SLR but also by changes in local exposure. The frequency of nuisance flooding along the US east coast is expected to accelerate further in the future (Sweet and Park, 2014). The loss of coral reef cover and mangrove forests have also been shown to increase damages from storm surge events (e.g., Beck et al. 2018). Cyclones also affect marine life, habitats and fishing. There is some evidence that fish may evacuate storm areas or be redistributed by storm waves and currents (FAO, 2018; Sainsbury et al. 2018). Other examples of damage to fisheries from cyclones and storm surges can be found in FAO (2018: Chapter V, Table 1).

With regard to property losses, according to most projections, increasing losses from more intense cyclones are not offset by a possible reduction in frequency (Handmer et al. 2012). While the relation between aggregate damages and frequency may be linear, the relationship between intensity and damages is most probably highly nonlinear; with research suggesting a 10% increase in wind speed associated with a 30–40% increase in damages (e.g., Strobl, 2012). Although it is clear that direct damages from cyclones could increase, investigations into the economic impact of past cyclone events is less common, as these are much more difficult to identify. Examples of such work include Strobl (2012) on hurricane impacts in the Caribbean, Haque and Jahan (2016) on TC Sidr in Bangladesh, Jakobsen (2012) on Hurricane Mitch in Nicaragua, and Taupo and Noy (2017) on TC Pam in Tuvalu. The relation between changes in TCs and property losses is complex, and there are indications that wind shear changes may have larger impact than changes in global temperatures (Wang and Toumi, 2016). With regard to loss of life, total fatalities and mortality from cyclone-related coastal flooding is globally declining, probably as a result of improved forecasting and evacuation, although in some low-income countries mortality is still high (Paul, 2009; Lumbroso et al. 2017; Bouwer and Jonkman, 2018). A global analysis finds that despite adaptation efforts, further SLR could increase storm surge mortality in many parts of the developing world (Lloyd et al. 2016).

An assessment of future changes in coastal impacts based on direct downscaling of indicators of flooding such as total water level and number of hours per year with breakwater overtopping over a given threshold for port operability is provided by Camus et al. (2017). These indicators are multivariable and include the combined effect of SLR, storm surge, astronomical tide and waves. Regional projected wave climate is downscaled from global multi-model projections from 30 CMIP5 model realisations. For example, projections by 2100 under the RCP8.5 scenario show a spatial variability along the coast of Chile with port operability loss between 600–800 h yr^{-1} and around 200 h yr^{-1} relative to present (1979–2005) conditions. Although wave changes are included in projected overtopping distributions, future changes of operability are mainly due to the SLR contribution.

6.3.3 Risk Management and Adaptation

The most effective risk management strategy in the last few decades has been the development of early warning systems for cyclones (Hallegatte, 2013). Generally, however, a lack of familiarity with the changed nature of storms prevails. Powerful storms often generate record storm surges (Needham et al. 2015), such as in the cases of Cyclone Nargis and Typhoon Haiyan but surge warnings had been less well understood and followed because they had tended to be new or rare to the locality (Lagmay et al. 2015). A US study on storm surge warnings highlights the issue of the right timing to warn, as well as the difficulty in delivering accurate surge maps (Morrow et al. 2015). Previous experience with warnings that were not followed by hazard events show the 'crying wolf' problem leading many to ignore future warnings (Bostrom et al. 2018).

There is scant literature on the management of storms that follow less common trajectories. The most recent and relatively well-studied ones are Superstorm Sandy in 2012 in the USA and Typhoon Haiyan in 2013 in the Philippines. These two storms were unexpected and having underestimated the levels of impact, people ignored warnings and evacuation directives. In the case of Typhoon Haiyan, the dissemination of warnings via scripted text messages were ineffective without an explanation of the difference between Haiyan's accompanying storm surge and that of other 'normal' storms to which people were used to (Lejano et al. 2016). Negative experiences of previous evacuations also lead to the reluctance of authorities to issue mandatory evacuation orders, for example, during Superstorm Sandy (Kulkarni et al. 2017), and contributes to a preventable high number of casualties (Dalisay and De Guzman, 2016). These examples also show that saving lives and assets through warning and evacuation is limited. Providing biophysical protection measures as well as improving self-reliance during such events can complement warning and evacuation.

After the storms, retreat or rebuild options exist. Rebuilding options can depend on whether insurance is still affordable after the event. Buyout programs, a form of 'managed retreat' whereby government agencies pay people affected by extreme weather events to relocate to safer areas, gained traction in recent years as a potential solution to reduce exposure to changing storm surge and flood risk. The decision to retreat or rebuild *in situ* depends, at least partially, on how communities have recovered in the past and therefore on the perceived success of a future recovery (Binder, 2014). However, political and jurisdictional conflicts between local, regional, and national government over land management responsibilities, lack of coordinated nation-wide adaptation plans, and clashes between individual and community needs have led to some unpopular buyout programs after Hurricane Sandy (Boet-Whitaker, 2017). Relocation (i.e., managed retreat) is often very controversial, can incur significant political risk even when it is in principle voluntary (Gibbs et al. 2016), and is rarely implemented with much success at larger scales (Beine and Parsons, 2015; Hino et al. 2017). In addition, managed retreats are often fraught with legal, distributional and human rights issues, as seen in the case of resettlements after Typhoon Haiyan (Thomas, 2015; see also Cross-Chapter Box 5 in Chapter 1), and extend to loss of cultural heritage and indigenous qualities in the case of small island states.

If rebuilding *in situ* is pursued after catastrophic events and without decreased exposure, it is often accompanied by actions that aim to reduce vulnerability in order to adapt to the increasing risk (Harman et al. 2013). In many cases, resilient designs and sustainable urban plans integrating climate change concerns, that are inclusive of vegetation barriers as coastal defences and hybrid designs, are considered (Cheong et al. 2013; Saleh and Weinstein, 2016). However, often more physical structures that are known to be less sustainable in the long-term, but potentially more protective in the short-term, are constructed (Knowlton and Rotkin-Ellman, 2014; Rosenzweig and Solecki, 2014). Anticipatory planning approaches are under way to warn and enable decision making in time (Bloemen et al. 2018; Lawrence et al. 2018).

6.4 Marine Heatwaves and their Implications

AR5 concluded that it is *virtually certain* that the global ocean temperature in the upper few hundred meters has increased from 1971–2010 (Rhein et al. 2013), and that the temperature is projected to further increase during the 21st century (Collins et al. 2013). For an update on observed and projected long-term changes in ocean temperature and heat, see Chapter 5.

Superimposed onto the long-term ocean warming trend are short-term extreme warming events, called MHWs, during which ocean temperatures are extremely high. Whereas the response of marine organisms and ecosystems to gradual trends in temperature has been assessed in AR5 (e.g., Hoegh-Guldberg et al. 2014; Pörtner et al. 2014), research on the response of the natural, physical and socioeconomic systems to MHWs has newly emerged since AR5. Notable exceptions are studies on the effect of MHWs on intertidal systems and tropical coral reef ecosystems, which have been already assessed in AR5 (Gattuso et al. 2014; Pörtner et al. 2014).

MHWs are periods of extremely high ocean temperatures that persist for days to months, can extend up to thousands of kilometres and can penetrate multiple hundreds of metres into the deep ocean (see SROCC Glossary; Hobday et al. 2016a; Scannell et al. 2016; Benthuysen et al. 2018). A MHW is an event at a particular place and time of the year that is rare and predominately, but not exclusively, defined with a relative threshold; that is, an event rarer than 90th or 99th percentile of a probability density function. By definition, the characteristics of what is called a MHW may therefore vary from place to place in an absolute sense. Different metrics are used to quantify changes in MHW characteristics, such as frequency, duration, intensity, spatial extent and severity. To monitor and predict coral bleaching risk, the metric degree heating week (DHW; e.g., Eakin et al. 2010) is often used, which combines the effect of duration and magnitude of the heatwave.

6.4.1 Observations and Key Processes, Detection and Attribution, Projections

6.4.1.1 Recent Documented MHWs and Key Driving Mechanisms

MHWs have been observed and documented in all ocean basins over the last two decades (Figure 6.3a, Figure 6.2, Table 6.2). Prominent examples include the Northeast Pacific 2013–2015 MHW (often called 'The Blob'; Bond et al. 2015), the Yellow Sea/East China Sea 2016 MHW (KMA, 2016; KMA, 2017; KMA, 2018), the Western Australia 2011 MHW (Pearce and Feng, 2013; Kataoka et al. 2014), and the Northwest Atlantic 2012 MHW (Mills et al. 2013).

The dominant ocean and/or atmospheric processes leading to the buildup, persistence and decay of MHWs vary greatly among the individual MHWs and depend on the location and time of occurrence. One of the most important global driver of MHWs are El Niño events (Oliver et al. 2018a). During El Niño events, the SST, in particular of the central and eastern equatorial Pacific and the Indian Ocean, are anomalously warm (see Section 6.5). MHWs may also be associated with other large-scale modes of climate variability, such as the Pacific Decadal Oscillation (PDO), AMO, Indian Ocean Dipole (IOD), North Pacific Oscillation and NAO, which modulate ocean temperatures at the regional scale (Benthuysen et al. 2014; Bond et al. 2015; Chen et al. 2015b; Di Lorenzo and Mantua, 2016). These modes can change the strength, direction and location of ocean currents that build up areas of extreme warm waters, or they can change the air-sea heat flux, leading to a warming of the ocean surface from the atmosphere. For example, predominant La Niña conditions in 2010 and 2011 strengthened and shifted the Leeuwin Current southward along the west coast of Australia leading to the Western Australia 2011 MHW (Pearce and Feng, 2013; Kataoka et al. 2014). Another example is The Blob, which emerged in 2013 in response to teleconnections between the North Pacific and the weak El Niño that drove strong positive sea level pressure anomalies across the northeast Pacific inducing a smaller heat loss from the ocean (Bond et al. 2015; Di Lorenzo and Mantua, 2016). Low sea ice concentrations in the Arctic, however, may have also played a role (Lee et al. 2015a).

The buildup and decay of extreme warm SSTs may also be caused by small-scale atmospheric and oceanic processes, such as ocean mesoscale eddies or local atmospheric weather patterns (Carrigan and Puotinen, 2014; Schlegel et al. 2017a; Schlegel et al. 2017b). For example, the Tasman Sea 2015–2016 MHW was caused by enhanced southward transport in the East Australian current driven by increased wind stress curl across the mid-latitude South Pacific (Oliver and Holbrook, 2014; Oliver et al. 2017) with local downwelling-favourable winds also having played a role in the subsurface intensification of the MHW (Schaeffer and Roughan, 2017). In addition, the 2016 MHW in the southern part of the Great Barrier Reef was mitigated by the ETC Winston that passed over Fiji on February 20th. The cyclone caused strong winds, cloud cover and rain, which lowered SST and prevented corals from bleaching (Hughes et al. 2017b).

6.4.1.2 Detection and Attribution of MHW Events

The upper ocean temperature has significantly increased in most regions over the last few decades, with anthropogenic forcing *very likely* being the main driver (Bindoff et al. 2013). Concurrent with the long-term increase in upper ocean temperatures, MHWs have become more frequent, extensive and intense (Frölicher and Laufkötter, 2018; Oliver et al. 2018a; Smale et al. 2019). Analysis of satellite daily SST data reveal that the number of MHW days exceeding the 99th percentile, calculated over the 1982–2016 period, has doubled globally between 1982 and 2016, from about 2.5 heatwave days yr^{-1} to 5 heatwave days yr^{-1} (Frölicher et al. 2018; Oliver et al. 2018a). At the same time, the maximum intensity of MHWs has increased by 0.15ºC and the spatial extent by 66% (Frölicher et al. 2018). Using a classification system to separate MHWs into categories (I-IV, depending on the level to which SSTs exceed local averages), Hobday et al. (2018) show that the occurrence of MHWs has increased for all categories over the past 35 years with the largest increase (24%) in strong (Category II) MHW events. In 2016, about a quarter of the surface ocean experienced either the longest or most intense MHW (Hobday et al. 2016a; Figure 6.3b).

The observed trend towards more frequent, intense and extensive MHWs, defined relative to a fixed baseline period, is *very likely* due to the long-term anthropogenic increase in mean ocean temperatures, and cannot be explained by natural climate variability (Frölicher et al. 2018; Oliver et al. 2018a; Oliver, 2019). As climate models project a long-term increase in ocean temperatures over the 21st century (Collins et al. 2013), a further increase in the probability of MHWs under continued global warming can be expected (see Section 6.4.1.3). Extending the analysis to the pre-satellite period (before 1982) by using a combination of daily *in situ* measurements and gridded monthly *in situ* based data

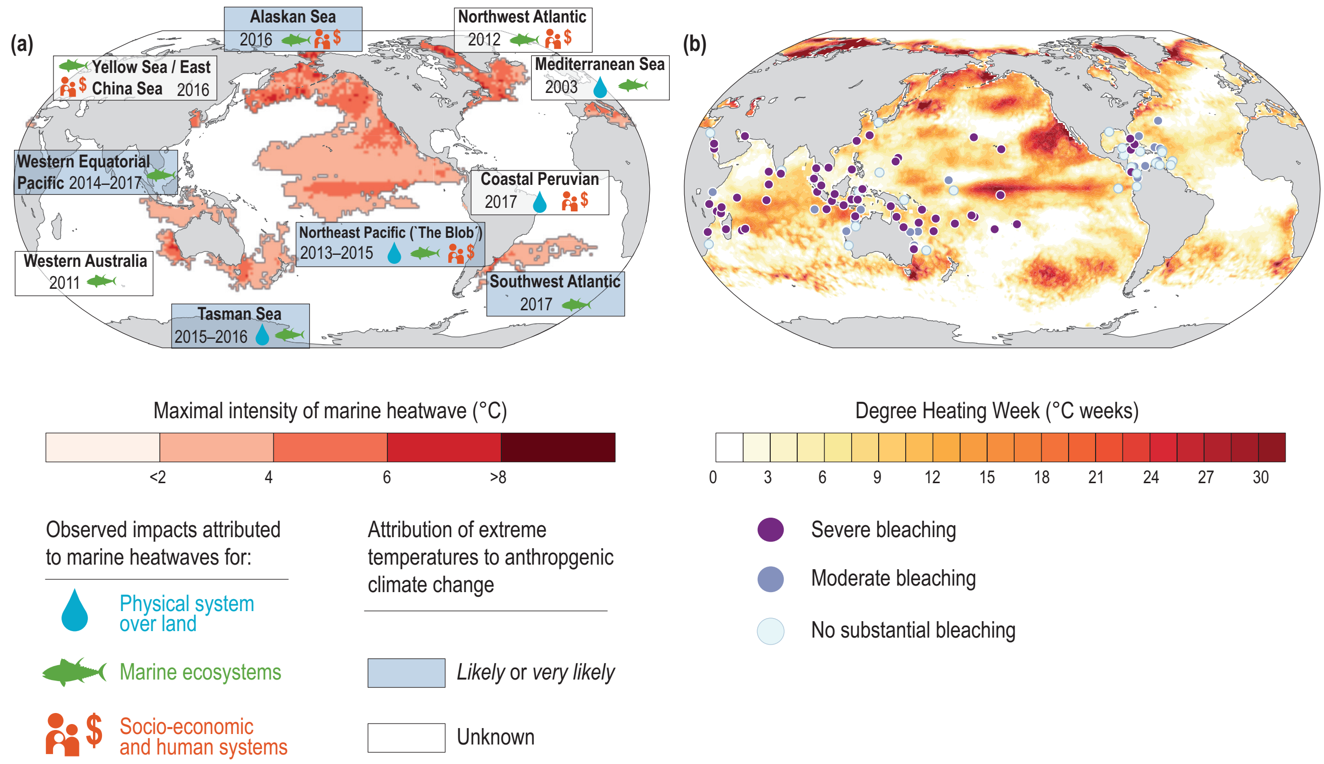

Figure 6.3 | Examples of recent marine heatwaves (MHWs) and their observed impacts. **(a)** Examples of documented MHWs over the last two decades and their impacts on natural, physical and socioeconomic systems. The colour map shows the maximum sea surface temperature (SST) anomaly during the MHW using the National Oceanic and Atmospheric Administration's (NOAA) daily Optimum Interpolation SST dataset (Reynolds et al. 2007; Banzon et al. 2016). A MHW is defined here as a set of spatially and temporally coherent grid points exceeding the 99th percentile. The 99th percentile is calculated over the 1982–2011 reference period after de-seasonalising the data. Red shading of the boxes indicates if the likelihood of MHW occurrence has increased due to anthropogenic climate change, and symbols denote observed impacts on physical systems over land, marine ecosystems, and socioeconomic and human systems. Figure is updated from Frölicher and Laufkötter (2018) and is not a complete compilation of all documented MHWs. **(b)** The record warming years 2015 and 2016 and the global extent of mass bleaching of corals during these years. The colour map shows the Degree Heating Week (DHW) annual maximum over 2015 and 2016 from NOAA's Coral Reef Watch Daily Global 5 km Satellite Coral Bleaching Heat Stress Monitoring Product Suite v.3.1 (Liu et al. 2014a). The DHW describes how much heat has accumulated in an area over the past twelve weeks by adding up any temperatures that exceed 1°C above the maximum summertime mean (e.g., Eakin et al. 2010). Symbols show reef locations that are assessed in Hughes et al. (2018a) and indicate where severe bleaching affected more than 30% of corals (purple circles), moderate bleaching affected less than 30% of corals (blue circles), and no substantial bleaching was recorded (light blue circles).

sets, Oliver et al. (2018a) show that the global frequency and duration of MHWs have increased since 1925. At regional scale, MHWs have become more common in 38% of the world's coastal ocean over the last few decades (Lima and Wethey, 2012). In tropical reef systems, the interval between recurrent MHWs and associated coral bleaching events has diminished steadily since 1980, from once every 25 to 30 years in early 1980s to once every 6 years in 2016 (Hughes et al. 2018a). Due to the scarcity of below surface temperature data with high temporal and spatial resolution, it is currently unknown if and how MHWs at depth have changed over the past decades.

Several attribution studies (summarised in Table 6.2) have investigated if the likelihood of individual MHW events has changed due to anthropogenic warming. On a global scale and at present day (2006–2015), climate models suggest that 84–90% (*very likely* range) of all globally occurring MHWs are attributable to the temperature increase since 1850–1900 (Fischer and Knutti, 2015; Frölicher et al. 2018). Attribution studies on individual MHW events show that the intensity of the western tropical Pacific MHW in 2014 (Weller et al. 2015), the intensity of the Alaskan Sea 2016 MHW (Oliver et al. 2018b; Walsh et al. 2018) and the extreme SSTs in the central equatorial Pacific in 2015–2016 can be fully attributed to anthropogenic warming. In other words, the aforementioned studies show that such events could not have occurred without the temperature increase since 1850–1900. In addition, extreme SSTs in the northeast Pacific in 2014 have become about five times more likely with human-induced global warming (Wang et al. 2014a; Kam et al. 2015; Weller et al. 2015). The Tasman Sea 2015–2016 MHW was 330 times (for duration) and 6.8 times (for intensity) more likely with anthropogenic climate change than without (Oliver et al. 2017), and the northern Australia 2016 MHW was up to fifty times more likely due to anthropogenic climate change (Weller et al. 2015; King et al. 2017; Lewis and Mallela, 2018; Newman et al. 2018; Oliver et al. 2018b). Also the risk of the Great Barrier Reef bleaching event in 2016 was increased due to anthropogenic climate change (King et al. 2017; Lewis and Mallela, 2018). Even though natural variability is still needed for the events to occur, these studies show that most of the individual MHW events analysed so far have a clear human-induced signal. However, such attribution studies have not been undertaken for all major individual MHW events yet (e.g., five out of ten MHWs indicated in Figure 6.3a have not been assessed), and it is therefore still unknown for some of the observed individual MHW events if they have an anthropogenic signal or not (labelled as 'unknown' in Figure 6.3a).

We conclude that it is *very likely* that MHWs have increased in frequency, duration and intensity since pre-industrial (1850–1900), and that between 2006–2015 most MHWs (84–90%; *very likely* range) are attributable to the temperature increase since 1850–1900. Only few studies on the attribution of individual MHW events exist, but they all point to human influence on recent MHW events.

6.4.1.3 Future Changes

MHWs will increase in frequency, duration, spatial extent and intensity throughout the ocean under future global warming (Oliver et al. 2017; Ramírez and Briones, 2017; Alexander et al. 2018; Frölicher et al. 2018; Frölicher and Laufkötter, 2018; Darmaraki et al. 2019). Projections based on 12 CMIP5 Earth system models suggest that, on global scale, the probability of MHWs exceeding the pre-industrial (1850–1900) 99th percentile will *very likely* increase by a factor of 20–27 by 2031–2050 and *very likely* by a factor of 46–55 by 2081–2100 under the RCP8.5 greenhouse gas (GHG) scenario (Figure 6.4a; Frölicher et al. 2018). In other words, a one-in-hundred-day event at pre-industrial levels is projected to become a one-in-four-day event by 2031–2050 and a one-in-two-day event by 2081–2100. The duration of MHW is projected to *very likely* increase from 8–10 days at 1850–1900, to 126–152 days in 2081–2100 under the RCP8.5 scenario (Frölicher et al. 2018). The maximum intensity (maximum exceedance of the 1850–1900 99th percentile) will *very likely* increase from 0.3ºC–0.4ºC in 1850–1900, to 3.1ºC–3.8ºC in 2081–2100 under the RCP8.5 scenario. Under the RCP2.6 scenario, the magnitude of changes in the different MHW metrics would be substantially reduced (Frölicher et al. 2018). For example, the probability ratio would *very likely* increase by a factor of 16–24 by 2081–2100 for RCP2.6; less than half of that is projected for the RCP8.5. The magnitude of changes in the probability ratio scales with global mean atmospheric surface temperature and is independent of the warming path (Figure 6.4b), that is, it does not depend on whether a particular warming level is reached sooner (RCP8.5) or later (RCP2.6).

The projected changes in MHWs will not be globally uniform. CMIP5 models project that the largest increases in the probability of MHWs will occur in the tropical ocean, especially in the western tropical Pacific, and the Arctic Ocean (Figure 6.4c,d), and that most of the large marine ecosystems will also experience large increases in the

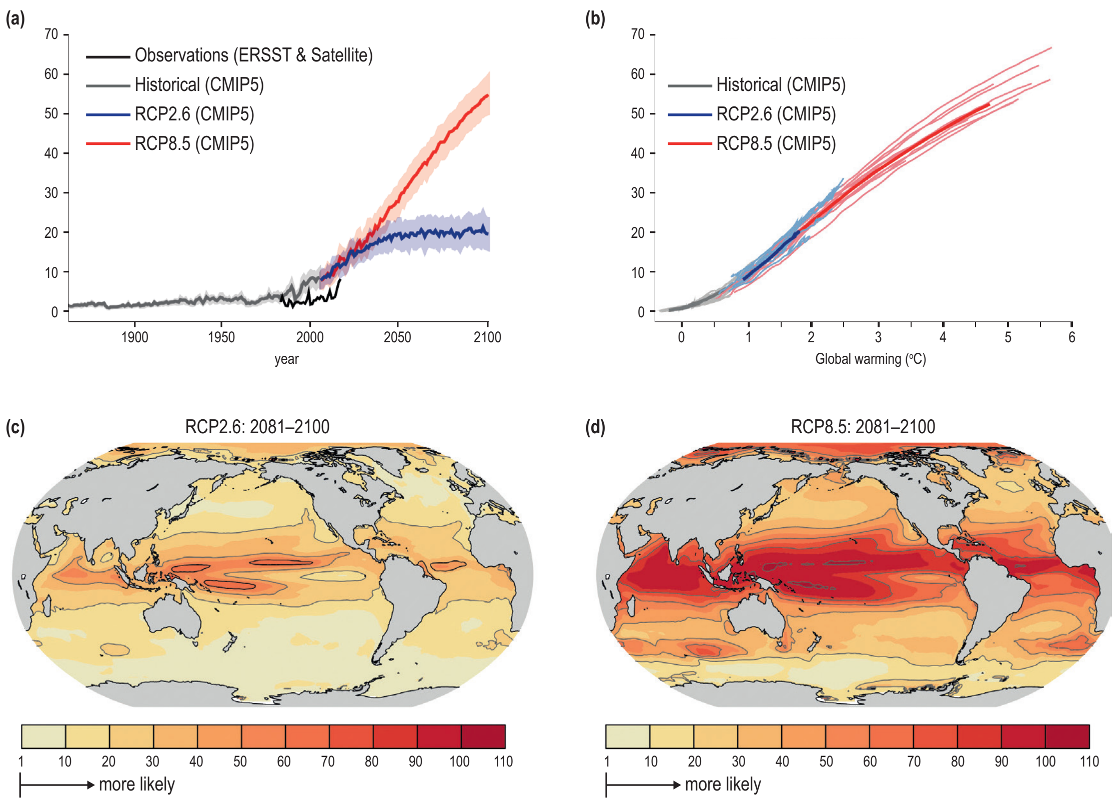

Figure 6.4 | Global and regional changes in the probability ratio of marine heatwaves (MHWs). The probability ratio is the fraction by which the number of MHW days yr^{-1} has changed since 1850–1900. **(a)** Changes in the annual mean probability ratio of MHWs exceeding the 99th percentile of pre-industrial local daily sea surface temperature (SST) averaged over the ocean. The thick lines represent the multi-model averages of 12 climate models that participated in the Coupled Model Intercomparison Project Phase 5 (CMIP5) covering the 1861–2100 period for the Representative Concentration Pathway (RCP) 8.5 and RCP2.6 scenarios, respectively. The shaded bands indicate the 90% confidence interval of the standard error of the mean. The black line shows an observational-based estimate. As daily SST data are available only for the 1982–2016 period, we assume that the observed mean temperature change is the main cause of the change in frequency of extremes (Frölicher et al. 2018; Oliver, 2019). We therefore subtracted first the differences between 1854–1900 and 1982–2016 obtained from the extended reconstructed SST Version 4 dataset (ERSSTv4; Huang et al. 2015a) from the daily satellite data before calculating the 99th percentile for the observations. **(b)** Same as (a), but the probability ratio is plotted for different levels of global surface atmospheric warming and for the individual models. The simulated time series in (b) are smoothed with a 10-year running mean. **(c,d)** Simulated regional changes in the multi-model mean probability ratio of MHWs exceeding the preindustrial 99th percentile in 2081–2100 for the (c) RCP2.6 scenario and the (d) RCP8.5 scenario. The grey contours in (c,d) highlight the spatial pattern. Figure is modified from Frölicher et al. (2018).

number of MHW days (Alexander et al. 2018; Frölicher et al. 2018). Smallest increases are projected for the Southern Ocean. In addition, MHW events in the Great Barrier Reef, such as the one associated with the bleaching in 2016, are projected to be at least twice as frequent under 2°C global warming than they are today (King et al. 2017). The magnitude of projected changes at the local scale is uncertain, partly due to issues of horizontal and vertical resolution of CMIP5-type Earth system models. Only a few studies have used higher resolution oceanic models (eddy-resolving) to assess the local-to-regional changes in MHW characteristics. For example, regional high-resolution coupled climate model simulations suggest that the Mediterranean Sea will experience at least one long lasting MHW every year by the end of the 21st century under the RCP8.5 scenario (Darmaraki et al. 2019), and eddy-resolving ocean model simulations project a further increase in the likelihood of extreme temperature events in the Tasman Sea (Oliver et al. 2014; Oliver et al. 2015; Oliver et al. 2017).

Most of the global changes in the probability of MHWs, when defined relative to a fixed temperature climatology and using coarse resolution CMIP5-type climate models, are driven by the global-scale shift in the mean ocean temperature (Alexander et al. 2018; Frölicher et al. 2018). However, previously ice-covered regions, such as the Arctic Ocean, will exhibit larger SST variability under future global warming. This is because of an enhanced SST increase in summer due to sea ice retreat, but SST remaining near the freezing point in winter (Carton et al. 2015; Alexander et al. 2018). When contrasting the changes in the probability of MHWs with land-based heatwaves (Fischer and Knutti, 2015), it is evident that MHWs are projected to occur more frequently (Frölicher et al. 2018; Frölicher and Laufkötter, 2018). This is because the temperature variability is much smaller in ocean surface waters than in the atmosphere (Frölicher and Laufkötter, 2018).

We conclude that there is *very high confidence* that MHWs will increase in frequency, duration, spatial extent and intensity in all ocean basins under future global warming, mainly because of an increase in mean ocean temperature. However, higher resolution models are needed to make robust projections at the local-to-regional scale.

6.4.2 Impacts on Natural, Physical and Human Systems

6.4.2.1 Impacts on Marine Organisms and Ecosystems

Temperature plays an essential role in the biology and ecology of marine organisms (e.g., Pörtner, 2002; Pörtner and Knust, 2007; Poloczanska et al. 2013; Hoegh-Guldberg et al. 2014), and therefore extreme high ocean temperature can have large impacts on marine ecosystems. Recent studies show that MHWs have strongly impacted marine organisms and ecosystem services in all ocean basins (Smale et al. 2019) over the last two decades. Impacts include coral bleaching and mortality (Hughes et al. 2017b; Hughes et al. 2018a; Hughes et al. 2018b), loss of seagrass and kelp forests (Smale et al. 2019), shifts in species range (Smale and Wernberg, 2013), and local (Wernberg et al. 2013; Wernberg et al. 2016) and potentially global extinctions of coral species (Brainard et al. 2011).

A growing number of studies have reported that MHWs negatively affect corals and coral reefs through bleaching, disease, and mortality (see Chapter 5 for an extensive discussion on coral reefs and coral bleaching). The recent (2014–2017) high ocean temperatures in the tropics and subtropics triggered a pan-tropical episode of unprecedented mass bleaching of corals (100s of km^2), the third global-scale event after 1997–1998 and 2010 (Heron et al. 2016; Eakin et al. 2017; Hughes et al. 2017b; Eakin et al. 2018; Hughes et al. 2018a). The heat stress during this event was sufficient to cause bleaching at 75% of global reefs (Hughes et al. 2018a; Figure 6.3b) and mortality at 30% (Eakin et al. 2017), much more than any previously documented global bleaching event. In some locations, many reefs bleached extensively for the first time on record, and over half of the reefs bleached multiple times during the three year event. However, there were distinct geographical variations in bleaching, mainly determined by the spatial pattern and magnitude of the MHW (Figure 6.3b). For example, bleaching was extensive and severe in the northern regions of the Great Barrier Reef, with 93% of the northern Australian Great Barrier Reef coral suffering bleaching in 2016, but impacts were moderate at the southern coral reefs of the Great Barrier Reef (Brainard et al. 2018; Stuart-Smith et al. 2018).

Apart from strong impacts on corals, recent MHWs have demonstrated their potential impacts on other marine ecosystems and ecosystems services (Ummenhofer and Meehl, 2017; Smale et al. 2019). Two of the best studied MHWs with extensive ecological implications are the Western Australia 2011 MHW and the Northeast Pacific 2013–2015 MHW. The Western Australia 2011 MHW resulted in a regime shift of the temperate reef ecosystem (Wernberg et al. 2013; Wernberg et al. 2016). The abundance of the dominant habitat-forming seaweeds *Scytohalia dorycara* and *Ecklonia radiata* became significantly reduced and *Ecklonia* kelp forest was replaced by small turf-forming algae with wide ranging impacts on associated sessile invertebrates and demersal fish. The sea grass *Amphibolis antarctica* in Shark Bay underwent defoliation after the MHW (Fraser et al. 2014), and together with the loss of other sea grass species, these lead to releases of 2–9 Tg CO_2 to the atmosphere during the subsequent three years after the MHW (Arias-Ortiz et al. 2018). In addition, coral bleaching and adverse impacts on invertebrate fisheries were documented (Depczynski et al. 2013; Caputi et al. 2016). The Northeast Pacific 2013–2015 MHW also caused extensive alterations to open ocean and coastal ecosystems (Cavole et al. 2016). Impacts included increased mortality events of sea birds (Jones et al. 2018), salmon and marine mammals (Cavole et al. 2016), very low ocean primary productivity (Whitney, 2015; Jacox et al. 2016), an increase in warm water copepod species (Di Lorenzo and Mantua, 2016) and novel species compositions (Peterson et al. 2017). In addition, a coast wide bloom of the toxigenic diatom *Pseudo-nitzschia* resulted in the largest ever recorded outbreak of domoic acid along the North American west coast (McCabe et al. 2016). Domoic acid was detected in many marine mammals, such as whales, dolphins, porpoises, seals and sea lions. The elevated toxins in commercially harvested fish and invertebrates resulted in prolonged and geographically extensive closure of razor clam and crab fisheries.

Other MHWs also demonstrated the vulnerability of marine organisms and ecosystems to extremely high ocean temperatures. The Northwest Atlantic 2012 MHW strongly impacted coastal ecosystems by causing a northward movement of warm water species and local migrations of some species (e.g., lobsters) earlier in the season (Mills et al. 2013; Pershing et al. 2015). The Mediterranean Sea 2003 MHW lead to mass mortalities of macro-invertebrate species (Garrabou et al. 2009) and the Tasman Sea 2015–2016 MHW had impacts on sessile, sedentary and cultured species in the shallow, near-shore environment including outbreaks of disease in commercially viable species (Oliver et al. 2017). *Vibrio* outbreaks were also observed in the Baltic Sea in response to elevated SSTs (Baker-Austin et al. 2013). The Alaskan Sea 2016 MHW favoured some phytoplankton species, leading to harmful algal blooms, shellfish poisoning events and mortality events in seabirds (Walsh et al. 2018; see chapter 3 for more details). Also, lower than average size of multiple groundfish species were observed including Pollock, Pacific cod, and Chinook salmon (Zador and Siddon, 2016). The Yellow Sea/East China Sea 2016 MHW killed a large number of different marine organisms in coastal and bay areas around South Korea (Kim and Han, 2017) and the Southwest Atlantic 2017 MHW lead to toxic algal blooms (Manta et al. 2018). The Coastal Peruvian 2017 MHW affected anchovies, which showed decreased fat content and early spawning as a reproductive strategy (IMPARPE, 2017), a behaviour usually seen during warm El Niño conditions (Ñiquen and Bouchon, 2004).

Based on the examples described above we conclude with *very high confidence* that a range of organisms and ecosystems have been impacted by MHWs across all ocean basins over the last two decades. Given that MHWs will *very likely* increase in intensity and frequency with further climate warming, we conclude with *high confidence* that this will push some marine organisms, fisheries and ecosystem beyond the limits of their resilience. These impacts will occur on top of those expected from a progressive shift in global mean ocean temperatures.

6.4.2.2 Impacts on the Physical System

MHWs can impact weather patterns over land via teleconnections causing drought, heavy precipitation or heat wave events. For example, the Northeast Pacific 2013–2015 MHW and the associated persistent atmospheric high-pressure ridge prevented normal winter storms from reaching the West Coast of the US and may have contributed to the drought conditions across the entire West Coast (Seager et al. 2015; Di Lorenzo and Mantua, 2016). The Tasman Sea 2015–2016 MHW has increased the intensity of rainfall that caused flooding in northeast Tasmania in January 2016 (see Box 6.1) and the Coastal Peruvian 2017 MHW caused heavy rainfall and flooding on the west coast of tropical South America (ENFEN, 2017; Echevin et al. 2018; Garreaud, 2018; Takahashi et al. 2018). Similarly, MHWs in the Mediterranean Sea may have amplified heatwaves (Feudale and Shukla, 2007; García-Herrera et al. 2010) and heavy precipitation events over central Europe (Messmer et al. 2017), as well as trigger intense ETCs over the Mediterranean Sea (González-Alemán et al. 2019). Such physical changes induced by MHWs may then also affect ecosystems and human systems on land (Reimer et al. 2015).

It should be noted that past and future impacts of MHWs on weather patterns over land depend not only on the duration and intensity of MHWs, but also on a wide range of different additional processes in the climate system such as the large-scale circulation of the atmosphere and oceans, and changes in the mean climate. Therefore, we conclude that there is currently *low confidence* in how MHWs impact the weather systems over land.

6.4.2.3 Impacts on the Human System

MHWs can also lead to significant socioeconomic ramifications when affecting aquaculture or important fishery species, or when triggering heavy rain or drought events on land. The Northwest Atlantic 2012 MHW, for example, had major economic impacts on the US lobster industry in 2015 (Mills et al. 2013). The MHWs lead to changes in lobster fishing practices and harvest patterns, because the lobsters moved from the deep offshore waters into shallower coastal areas much earlier in the season than usual causing a rapid rise in lobster catch rates. Together with a supply chain bottleneck, the record catch outstripped market demand and contributed to a collapse in lobster prices (Mills et al. 2013). Even though high catch volumes were reported, the price collapse threatened the economic viability of many US and Canadian lobster fisheries. Economic impacts through changes in fisheries were also reported during the Northeast Pacific 2013–2015 MHW and the Alaskan Sea 2016 MHW. The Northeast Pacific 2013–2015 MHW led to closing of both commercial and recreational fisheries resulting in millions of USD in losses among fishing industries (Cavole et al. 2016). In addition, the toxin produced by the harmful algal blooms can be transferred through the marine food web and humans who eat contaminated fish, shellfish or crustaceans (Berdalet et al. 2016; Du et al. 2016; McCabe et al. 2016). The ingestion of such contaminated seafood products, the inhalation of aerosolised toxins or the skin contact with toxin-contaminated water may cause toxicity in humans. Symptoms in human associated with the ingestion of the contaminated seafood range from mild gastrointestinal distress to seizures, coma, permanent short-term memory loss and death (Perl et al. 1990). The ecological changes associated with the Alaskan Sea 2016 MHW impacted subsistence and commercial activities. For example, ice-based harvesting of seals, crabs and fish in western Alaska was delayed due to the lack of winter sea ice. MHWs can also impact the socioeconomic and human system through changes to weather patterns. For example, heavy rain associated with the Coastal Peruvian 2017 MHW triggered numerous landslides and flooding, which resulted in a death toll of several hundred, and widespread damage to infrastructure and civil works (United Nations, 2017).

Studies on the impact of MHWs on human systems are still relatively scarce, even though many show negative impacts on human health and economy. We therefore conclude with *medium confidence* that MHWs can negatively impact human health and economy.

6.4.3 Risk Management and Adaptation, Monitoring and Early Warning Systems

Risk management strategies to respond to MHWs include early warning systems as well as seasonal (weeks to several months) and multi-annual predictions systems. Since 1997, the National Oceanic and Atmospheric Administration's (NOAA) Coral Reef Watch has used satellite SST data to provide near real-time warning of coral bleaching (Liu et al. 2014a). These satellite-based products, along with NOAA Coral Reef Watch's four month coral bleaching outlook based on operational climate forecast models (Liu et al. 2018), and coral disease outbreak risk (Heron et al. 2010) provide critical guidance to coral reef managers, scientists, and other stakeholders (Tommasi et al. 2017b; Eakin et al. 2018). These products are also used to implement proactive bleaching response plans (Rosinski et al. 2017), brief stakeholders, and allocate monitoring resources in advance of bleaching events, such as the 2014–2017 global coral bleaching event (Eakin et al. 2017). For example, Thailand closed ten reefs for diving in advance of the bleaching peak in 2016, while Hawaii immediately began preparation of resources both to monitor the 2015 bleaching and to place specimens of rare corals in climate controlled, onshore nurseries in response to these forecast systems (Tommasi et al. 2017b). New measurement techniques, such as Argo and deep Argo floats, may help to further develop prediction systems for subsurface MHWs, but such systems are not yet in place.

SST forecasts ranging from seasonal to decadal (5–10 years) have also been used or are planned to be used as early warning systems for multiple other ecosystems and fisheries in addition to coral reefs, including aquaculture, lobster, sardine, and tuna fisheries (Hobday et al. 2016b; Payne et al. 2017; Tommasi et al. 2017b). For example, seasonal forecasts of SST around Tasmania may help farm managers of salmon aquaculture to prepare and respond to upcoming MHWs by changing stocking densities, varying feed mixes, transferring fish to different locations in the farming region and implementing disease management (Spillman and Hobday, 2014; Hobday et al. 2016b). Skilful multi-annual to decadal SST predictions may also inform and improve decisions about spatial and industrial planning, as well as the management of various extractive sectors such as the adjustments to quotas for internationally shared fish stocks (Tommasi et al. 2017a). It has been shown that global climate forecasts have significant skill in predicting the occurrence of above average warm or cold SST events at decadal timescales in coastal areas (Tommasi et al. 2017a), but barriers to their widespread usage in fishery and aquaculture industry still exist (Tommasi et al. 2017b).

Even with a monitoring and prediction system in place, MHWs have developed without warning and had catastrophic effects (Payne et al. 2017). For example, governmental agencies, socioeconomic sectors, public health officials and citizens were not forewarned of the Coastal Peruvian 2017 MHW, despite a basin-wide monitoring system across the Pacific. The reason was partly due to a coastal El Niño definition problem and a new government (in Nicaragua) that may have hindered actions (Ramírez and Briones, 2017). Therefore, early warning systems should not only provide predictions of physical changes, but should also connect different institutions to assist decision makers in performing time-adaptive measures (Chang et al. 2013).

Monitoring and prediction systems are important and can be advanced by the use of common metrics to describe MHWs. So far, MHWs are often defined differently in the literature, and it is only recently that a categorising scheme (Categories I to IV; based on the degree to which temperatures exceed the local climatology), similar to what is used for hurricanes, has been developed (Hobday et al. 2018). Such a categorising scheme, can easily be applied to real data and predictions, and may facilitate comparison, public communication and familiarity with MHWs. Similar metrics (e.g., DHW) have been successfully developed and used to identify ocean regions where conditions conducive to coral bleaching are developing.

6.5 Extreme ENSO Events and Other Modes of Interannual Climate Variability

6.5.1 Key Processes and Feedbacks, Observations, Detection and Attribution, Projections

6.5.1.1 Extreme El Niño, La Niña

AR5 (Christensen et al. 2013) and SREX do not provide a definition for an extreme El Niño but mention such events, especially in the context of the 1997–1998 El Niño and its impacts. AR5 and SREX concluded that confidence in any specific change in ENSO variability in the 21st century is low. However, they did note that due to increased moisture availability, precipitation variability associated with ENSO is likely to intensify. Since AR5 and SREX, there is now a limited body of literature that examines the impact of climate change on ENSO over the historical period.

Palaeo-ENSO studies suggest that ENSO was highly variable throughout the Holocene, with no evidence for a systematic trend in ENSO variance (Cobb et al. 2013) but with some indication that the ENSO variance over 1979–2009 has been much larger than that over 1590–1880 (McGregor et al. 2013). Palaeo-ENSO reconstruction for the past eight centuries suggests that central Pacific ENSO activity has increased between the last two decades (1980–2015; Liu et al. 2017b), with an increasing number of central Pacific El Niño events compared to east Pacific El Niño events (Freund et al. 2019). Further proxy evidence exists for changes in the mean state of the equatorial Pacific in the last 2000 years (Rustic et al. 2015; Henke et al. 2017). Simulations using an Earth System Model indicate significantly higher ENSO variance during 1645–1715 than during the 21st century warm period, though it is unclear whether these simulated changes are realistic (Keller et al. 2015). For the 20th century, the frequency and intensity of El Niño events were high during 1951–2000, in comparison with the 1901–1950 period (Lee and McPhaden, 2010; Kim et al. 2014b; Roxy et al. 2014). Current instrumental observational records are not long enough and the quality of data before 1950 is limited, to assert these changes with *high confidence* (Wittenberg, 2009; Stevenson et al. 2010) though the palaeo records mentioned here signal the emergence of a statistically significant increase in ENSO variance in recent decades.

Since SREX and AR5, an extreme El Niño event occurred in 2015–2016. This has resulted in significant new literature regarding physical processes and impacts but there are no firm conclusions regarding the impact of climate change on the event. The SST anomaly peaked toward the central equatorial Pacific causing floods in many regions of the world such as those in the west coasts of the USA and other parts of North America, some parts of South America close to Argentina and Uruguay, the UK and China (Ward et al. 2014; Ward et al. 2016; Zhai et al. 2016; Scaife et al. 2017; Whan and Zwiers, 2017; Sun and Miao, 2018; Yuan et al. 2018).

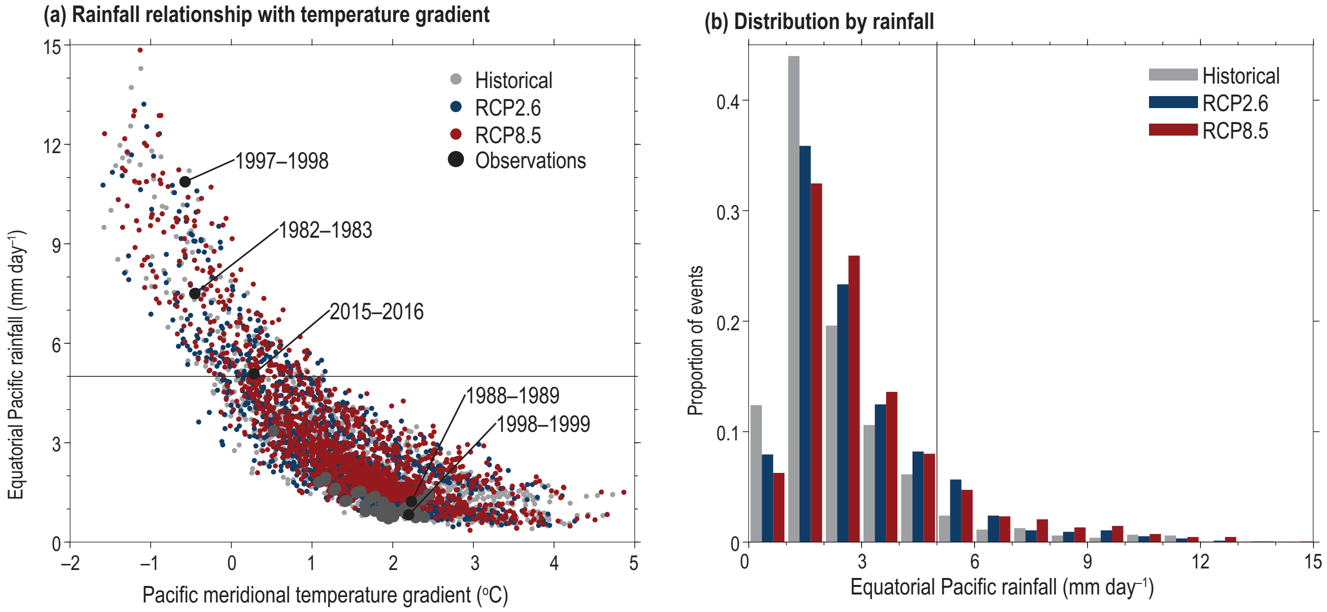

Figure 6.5 | Frequency of extreme El Niño Southern Oscillation (ENSO) events, adapted from Cai et al. (2014a). **(a)** December to February mean meridional sea surface temperature (SST) gradient (x-axis: 5ºN–10ºN, 210ºE–270ºE minus 2.5ºS–2.5ºN, 210ºE–270ºE) versus equatorial Pacific anomalous rainfall (y-axis: 5ºS–5ºN, 210ºE–270ºE). Data from only those Coupled Model Intercomparison Project Phase 5 (CMIP5) models that capture the observed relationship between Pacific SST and rainfall are shown. Black dots are from observations with extreme El Niño and extreme La Niña years indicated. The horizontal line denotes the threshold of 5 mm day^{-1} for an extreme event. **(b)** Histogram showing the relative frequency of rainfall rates. The vertical line denotes the 5 mm day^{-1} threshold. Higher counts of extreme events under the Representative Concentration Pathway (RCP)8.5 scenario suggest an increase in the frequency of extreme El Niño under global warming.

The main new body of literature concerns future projections of the frequency of occurrence and variability of extreme ENSO events with improved confidence (Cai et al. 2014a; Cai et al. 2018). These studies define extreme El Niño events as those El Niño events which are characterised by a pronounced eastward extension of the west Pacific warm pool and development of atmospheric convection, and hence a rainfall increase of greater than 5 mm day^{-1} during December to February (above 90th percentile), in the usually cold and dry equatorial eastern Pacific (Niño 3 region, 150ºW–90ºW, 5ºS–5ºN; Cai et al. 2014a), such as the 1982–1983, 1997–1998 and 2015–2016 El Niños (Santoso et al. 2017; Figure 6.5).

The background long-term warming puts the 2015–2016 El Niño among the three warmest in the instrumental records (24 El Niño events occurred during 1900–2018; Huang et al. 2016; Santoso et al. 2017). The 2015–2016 event can be viewed as the first emergence of an extreme El Niño in the 21st century – one which satisfies the rainfall threshold definition, but not characterised by the eastward extension of the west Pacific warm pool (L'Heureux et al. 2017; Santoso et al. 2017).

Based on the precipitation threshold, extreme El Niño frequency is projected to increase with the global mean temperatures (*medium confidence*) with a doubling in the 21st century under 1.5ºC of global warming, from about one event every 20 years during 1891–1990, to one every 10 years (Cai et al. 2014a; Figure 6.5). The increase in frequency continues for up to a century even after global mean temperature has stabilised at 1.5ºC, thereby challenging the limits to adaptation, and hence indicates high risk even at the 1.5ºC threshold (Wang et al. 2017; Hoegh-Guldberg et al. 2018). Meanwhile, the La Niña events also tend to increase in frequency and double under RCP8.5 (Cai et al. 2015), but indicate no further significant changes after global mean temperatures have stabilised (Wang et al. 2017). Particularly concerning is that swings from extreme El Niño to extreme La Niña (opposite of extreme El Niño) have been projected to occur more frequently under greenhouse warming (Cai et al. 2015). The increasing ratio of Central Pacific El Niño events to East Pacific El Niño events is projected to continue, under increasing emissions (Freund et al. 2019). Further, CMIP5 models indicate that the risk of major rainfall disruptions has already increased for countries where the rainfall variability is linked to ENSO variability. This risk will remain elevated for the entire 21st century, even if substantial reductions in global GHG emissions are made (*medium confidence*). The increase in disruption risk is caused by anthropogenic warming that drives an increase in the frequency and magnitude of ENSO events and also by changes in background SST patterns (Power et al. 2013; Chung et al. 2014; Huang and Xie, 2015). While many of these studies have adopted the precipitation view of an extreme El Nino, studies also indicate an increase in SST variability for events with their main SST anomalies in the east Pacific (Cai et al. 2018). Also, a role of cross-equatorial winds has been identified (Hu and Fedorov, 2018).

6.5.1.2 Indian Ocean Basin-wide Warming and Changes in Indian Ocean Dipole (IOD) Events

The Indian Ocean has experienced consistent warming from the surface to 2,000 m during 1960–2015, with most of the warming occurring in the upper 300 m (Cheng et al. 2015; Nieves et al. 2015; Cheng et al. 2017; Gnanaseelan et al. 2017). New historical ocean heat content (OHC) estimates show an abrupt increase in the Indian Ocean upper 700 m OHC after 1998, contributing to more than 21% of the global ocean heat gain, despite representing only about 12%

of the global ocean area (Cheng et al. 2017; Makarim et al. 2019). The tropical Indian Ocean SST has warmed by 1.04ºC during 1950–2015, while the tropical SST warming is 0.83ºC and the global SST warning is 0.65ºC. More than 90% of the surface warming in the Indian Ocean has been attributed to changes in GHG emissions (Dong et al. 2014), with the heat redistributed in the basin via local ocean and atmospheric dynamics (Liu et al. 2015b), the ITF (Section 6.6.1; Susanto et al. 2012; Sprintall and Revelard, 2014; Lee et al. 2015b; Susanto and Song, 2015; Zhang et al. 2018) and the Walker circulation (Roxy et al. 2014; Abish et al. 2018).

The dynamic processes related to the projected changes in IOD under global warming have a large inter-model spread (Cai et al. 2013). The frequency of extreme positive IOD events are projected to increase by almost a factor of three, from a one-in-seventeen-year event in the 20th century to a one-in-six-year event in the 21st century (*low confidence*). The bias in the CMIP5 models and internal variability could enlarge the projected increase in the extreme positive IOD events (Li et al. 2016a; Hui and Zheng, 2018). The increase in IOD events is not linked to the change in the frequency of El Niño events but instead to mean state change—with weakening of both equatorial westerly winds and eastward oceanic currents in association with a faster warming in the western than the eastern equatorial Indian Ocean (Cai et al. 2014b). A combination of extreme ENSO and IOD events has led to a northward shift in the Intertropical Convergence Zone (ITCZ) during 1979–2015, which is expected to increase further in the future (Freitas et al. 2017).

6.5.2 Impacts on Human and Natural Systems

Increasing frequency of extreme ENSO and IOD events have the potential to have widespread impacts on natural and human systems in many parts of the globe. Though the occurrence of the extreme 2015–2016 El Niño has produced a large body of literature, it is still not clear how climate change may have altered such an impact, nor how such impacts might change in the future with increasing frequency of extreme ENSO events. We highlight here some studies that have attempted to assess the joint impact of mean change and variability. In addition to observed high variability of rainfall, severe weather events and impacts on TCs activity (Yonekura and Hall, 2014; Zhang and Guan, 2014; Wang and Liu, 2016; Zhan, 2017), extreme El Nino events have substantial impacts on natural systems which include those on marine ecosystems (Sanseverino et al. 2016; Mogollon and Calil, 2017; Ohman et al. 2017), such as severe and repeated bleaching of corals (Hughes et al. 2017a; Hughes et al. 2017b; Eakin et al. 2018), and glacial growth and retreat (Thompson et al. 2017). On the other hand, impacts on human, including managed systems are: increased incidences of forest fires (Christidis et al. 2018b; Tett et al. 2018), degraded air quality (Koplitz et al. 2015; Chang et al. 2016; Zhai et al. 2016) such as the dense haze over most parts of Indonesia and the neighboring countries in Southeast Asia as a result of prolonged Indonesian wildfires, thus imposing adverse impacts on public health in the affected areas (Koplitz et al. 2015; WMO, 2016), decreased agricultural yields in many parts of the globe (e.g., in most of the Pacific Islands countries, Thailand, eastern and southern Africa and others which resulted food insecurity, particularly in eastern and southern Africa (UNSCAP, 2015; WMO, 2016; Christidis et al. 2018b; Funk et al. 2018), and regional uptick in the number of reported cases of plague and hantavirus in Colorado and New Mexico, cholera in Tanzania, dengue in Brazil and Southeast Asia (Anyamba et al. 2019) and Zika virus in South America (Caminade et al. 2017), including increases in heat stroke cases (Christidis et al. 2018b). Substantial economic losses had resulted from droughts and floods across various parts of the globe due to teleconnections. For instance, direct losses of 10 billion USD (Sun and Miao, 2018; Yuan et al. 2018) and 6.5 billion USD (Christidis et al. 2018b) were estimated to have been incurred from severe urban inundation in cities along the Yangtze River in China and the extreme drought in Thailand, respectively.

ENSO events affect TCs activity through variations in the low-level wind anomalies, vertical wind shear, mid-level relative humidity, steering flow, the monsoon trough and the western Pacific subtropical high in Asia (Yonekura and Hall, 2014; Zhang and Guan, 2014). The subsurface heat discharge due to El Niño can intensify TCs in the eastern Pacific (Jin et al. 2014; Moon et al. 2015b). TCs are projected to become more frequent (~20–40%) during future-climate El Niño events compared with present climate El Niño events (*medium confidence*), and less frequent during future-climate La Niña events, around a group of small island nations (for example, Fiji, Vanuatu, Marshall Islands and Hawaii) in the Pacific (Chand et al. 2017). The Indian Ocean basin-wide warming has led to an increase in TC heat potential in the Indian Ocean over the last 30 years, however the link to the changes in the frequency of TCs is not robust (Rajeevan et al. 2013).

During the early stages of an extreme El Niño event (2015–2016 El Niño), there is an initial decrease in atmospheric CO_2 concentrations over the tropical Pacific Ocean, due to suppression of equatorial upwelling, reducing the supply of CO_2 to the surface (Chatterjee et al. 2017), followed by a rise in atmospheric CO_2 concentrations due reduced terrestrial CO_2 uptake and increased fire emissions (Bastos et al. 2018). It is not clear how a future increase in the frequency extreme events would modulate the carbon cycle on longer decadal time scales.

Studies on projections of changes in ENSO impacts or teleconnections are rather limited. Nevertheless, Power and Delage (2018) provide a multi-model assessment of CMIP5 models and their simulated changes in the precipitation response to El Niño in the future (Figure 6.6). They identify different combinations of changes that might further impact natural and human systems. El Niño causes either positive or negative precipitation anomalies in diverse regions of the globe. Dry El Niño teleconnection anomalies may be further strengthened by, either mean climate drying in the region (Amazon, Central America and Australia in June to August (JJA)), or a strengthening of the El Niño dry teleconnection, or both. Conversely, wet El Niño teleconnections can be further strengthened by either increases in mean precipitation (East Africa and southeastern South America in December to February (DJF)) or a strengthening of the El Niño wet teleconnection (southeastern South America in JJA), or both (Tibetan Plateau, DJF). However, a present day dry El Niño response may be dampened by a wet mean response (South, East and Southeast Asia in JJA) or a wet present day El Niño response may be weakened by a dry mean change (Southern Europe/Mediterranean

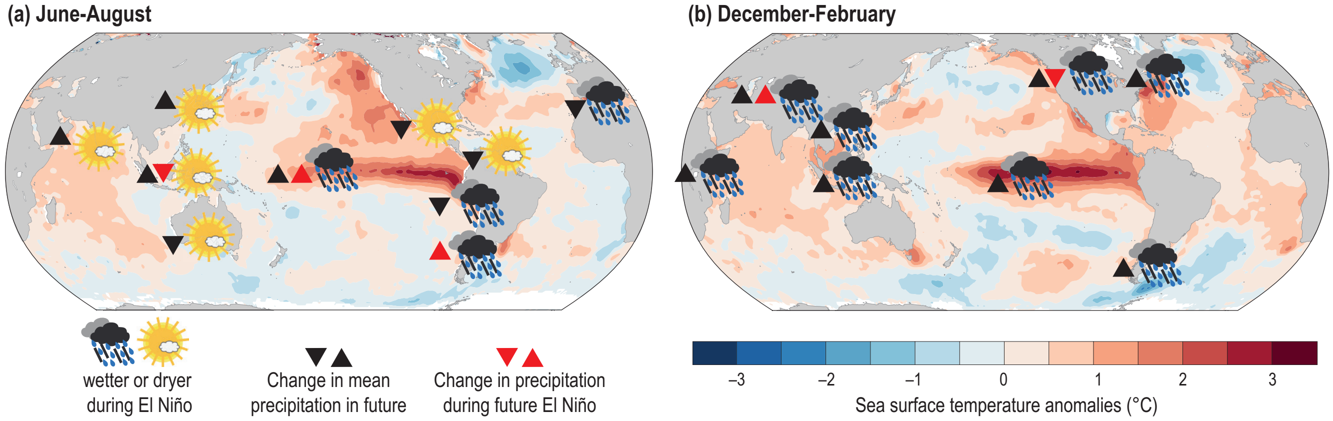

Figure 6.6 | Schematic figure indicating future changes in El Niño teleconnections based on the study of Power and Delage (2018). The background pattern of sea surface temperature (SST) anomalies (°C) are averaged from June 2015 to August 2015 (panel a) and December 2015 to February 2016 (panel b), during the most recent extreme El Niño event (anomalies computed with respect to 1986–2005). Symbols indicate present day teleconnections for El Niño events. Black arrows indicate if there is a model consensus on change in mean rainfall in the region. Red arrows indicate if there is a model consensus on change in the rainfall anomaly under a future El Niño event. Direction of the arrow indicates whether the response in precipitation is increasing (up) or decreasing (down). Significance is determined when two-thirds or more of the models agree on the sign.

and West Coast South America in JJA). Finally, changes in the mean and El Niño response may be in the opposite direction (Southeast Asia, JJA and Central North America, DJF). Such changes could have an impact on phenomena such as wildfires (Fasullo et al. 2018). However, in many other regions that are currently impacted by El Niño, e.g., regions of South America, studies have found no significant changes in the ENSO-precipitation relationship (Tedeschi and Collins, 2017) and agreement between models for many regions suggests *low confidence* in projections of teleconnection changes (Yeh et al. 2018).

Along with extreme El Niño events, abrupt warming in the Indian Ocean and extreme IOD events have largely altered the Asian and African monsoon, impacting the food and water security over these regions. As a response to rising global SSTs and partially due to extreme El Niño events, the NH summer monsoon showed substantial intensification during 1979–2011, with an increase in rainfall by 9.5% per degree Celsius of global warming (Wang et al. 2013). However, the Indian summer monsoon circulation and rainfall exhibits a statistically significant weakening since the 1950s. This weakening has been hypothesised to be a response to the Indian Ocean basin-wide warming (Mishra et al. 2012; Roxy et al. 2015) and also to increased aerosol emissions (Guo et al. 2016) and changes in land use (Paul et al. 2016). Warming in the north Indian Ocean has resulted in increasing fluctuations in the southwest monsoon winds and a three-fold increase in extreme rainfall events across central India (Roxy et al. 2017). The frequency and duration of heatwaves have increased over the Indian subcontinent, and these events are associated with the Indian Ocean basin-wide warming and frequent El Niños (Rohini et al. 2016). In April 2016, as a response to the extreme El Niño, Southeast Asia experienced surface air temperatures that surpassed national records, increased energy consumption, disrupted agriculture and resulted in severe human discomfort (Thirumalai et al. 2017). A strong negative IOD event in 2016 led to large climate impact on East African rainfall, with some regions recording below 50% of normal rainfall, leading to devastating drought, food insecurity and unsafe drinking water for over 15 million people in Somalia, Ethiopia and Kenya.

6.5.3 Risk Management and Adaptation

Risk management of ENSO events has focussed on two main aspects: better prediction and early warning systems, and better mechanisms for reducing risks to agriculture, infrastructure, fisheries and aquaculture, wildfire and flood management. Extreme ENSO events are rare, with three such events since 1950 and they are difficult to predict due to the different drivers influencing them (Puy et al. 2017). The impacts of ENSO events also vary between events and between the different regions affected (Murphy et al. 2014; Fasullo et al. 2018; Power and Delage, 2018) however, there is limited literature on the change in the impacts of extreme ENSO compared to other ENSO events. In addition, there are also no specific risk management and adaptation strategies for human and natural systems for more extreme events other than what is in place for ENSO events (see also Chapter 4, Section 4.4 for the response to sea level change, an observed impact of ENSO). A first step in risk management and adaptation is thus to better understand the impacts these events have and to identify conditions that herald such extreme events that could be used to better predict extreme ENSO events.

Monitoring and forecasting are the most developed ways to manage extreme ENSOs. Several systems are already in place for monitoring and predicting seasonal climate variability and ENSO occurrence. However, the sustainability of the observing system is challenging and currently the Tropical Pacific Observing System 2020 (TPOS 2020) has the task of redesigning such a system, with ENSO prediction as one of its main objectives. These systems could be further elaborated to include extreme ENSO events. Westerly wind events in the Western Tropical Pacific, (Lengaigne et al. 2004; Chen et al. 2015a; Fedorov et al. 2015) strong easterly wind events in the tropical Pacific (Hu and Fedorov, 2016; Puy et al. 2017), nonlinear interaction between air-sea fluxes and atmospheric deep convection (Bellenger et al. 2014; Takahashi and Dewitte, 2016) and advection of mean temperature by anomalous eastward zonal currents (Kim and Cai, 2014) are some of the factors that play an important role in the evolution of extreme ENSO events, which can be considered while improving the monitoring and forecasting system.

Despite the specificity of each extreme El Niño event, their forecasting is expected to improve through monitoring of recently identified precursory signals that peak in a window of two years before the event (Varotsos et al. 2016). An early warning system for coral bleaching associated, among other stressors, with extreme ENSO heat stress is provided by the NOAA Coral Reef Watch service with a 5 km resolution (Liu et al. 2018). The impacts of ENSO-associated extreme heat stress are heterogeneous, indicating the influence of other factors either biotic such as coral species composition, local adaptation by coral taxa reef depth or abiotic such as local upwelling or thermal anomalies (Claar et al. 2018). When identified and quantified, these factors can be used for risk analysis and risk management for these ecosystems.

In principle, it is easier to transfer the financial risk associated with extreme ENSO events through, for example, insurance products or other risk transfer instruments such as Catastrophe Bonds, than for more moderate events. An accurate prediction system is not required, but the measurement of these events, and quantification of likely impacts is required. As in other types of insurance systems, this can be done through, for example, calculations of average annual losses associated with extreme ENSO, and the design of appropriate financial instruments. Examples of research that can support the design of risk transfer instruments include Anderson et al. (2018) and Gelcer et al. (2018) for specific crops yields, and Aguilera et al. (2018) and Broad et al. (2002) for specific fisheries. Several risk transfer instruments have been implemented to deal with ENSO impacts, including parametric insurance based on SSTs for heavy rainfall damages, and another scheme for agricultural damages, both in Peru. Other examples include forecast-based financial aid (Red Cross Climate Centre, 2016). More broadly, other forms of risk management and governance can be designed with better information about the likely impacts of extreme ENSO events (e.g., Vignola et al. 2018).

6.6 Inter-Ocean Exchanges and Global Change

Section 3.6.5.1 in AR5 briefly described the Indonesian Throughflow (ITF) but did not explain its variability and impacts. Palaeoclimate record, observations, and climate model studies suggest that ITF plays an integral role in global ocean circulation, directly impacting mass, heat and freshwater budgets of the Pacific and Indian Oceans (*high confidence*). ITF is influenced by equatorial Pacific trade wind system which experienced an unprecedented intensification during 2001–2014, resulting in enhanced ocean heat transport from the Pacific to the Indian Ocean and influencing the rate of global temperature change (*medium confidence*). Yet, numerical models are not able to simulate the magnitude of decadal variability and the inter-ocean link, which means there is *low confidence* in future projections of the trade wind system.

6.6.1 Key Processes and Feedbacks, Observations, Detection and Attribution, Projections

In the last two decades, total water transport from the Pacific to the Indian Ocean and the Indian Ocean to the Atlantic Ocean has increased (*high confidence*). Increased ITF has been attributed to Pacific cooling and basin-wide warming in the Indian Ocean. The ITF annual average is 15 x 10^6 m^3 s^{-1} (Susanto et al. 2012). ITF varies from intraseasonal to decadal time scales. On seasonal time scale, South China Sea Throughflow controls freshwater flux and modulates the main ITF (Fang et al. 2010; Susanto et al. 2013; Lee et al. 2019; Wang et al. 2019; Wei et al. 2019). During the extreme El Niño of 1997–1998, the ITF transport was reduced to 9.2 x 10^6 m^3 s^{-1}. Based on observations and proxy records from satellite altimetry and gravimetry, in the last two decades, 1992–2012, ITF has been stronger (Sprintall and Revelard, 2014; Liu et al. 2015a; Susanto and Song, 2015), which translates to an increase in ocean heat-flux into the Indian Ocean (Lee et al. 2015b). Exchanges of heat and fresh water between ocean basins are important at the global scale (Flato et al. 2013). ITF may have played a key role in the slowdown of the Pacific SST warming during 1998–2013, and the rapid warming in the surface and subsurface Indian Ocean during this period (Section 6.5.1.2; Makarim et al. 2019), by transferring warm water from the western Pacific into the Indian Ocean (Lee et al. 2015b; Dong and McPhaden, 2018).

Under 1.5ºC warming both El Niño and La Niña frequencies may increase (see Section 6.5) and hence ITF variability may also increase. ITF is also influenced by the IOD events, with an increase in transport during a positive IOD and vice-versa during a negative IOD event (Potemra and Schneider, 2007; Pujiana et al. 2019). Positive IODs are projected to increase threefold in the 21st century as a response to changes in the mean state rather than changes in the El Niño frequency (Section 6.5.1.2; Cai et al. 2014b) and this may have an impact on the ITF, additional to the changes due to increasing extreme ENSO events. In response to greenhouse warming, climate models predict that on interannual time scale, it is *likely* that the mean ITF may decrease due to wind variability (Sen Gupta et al. 2016), but recent observation trend tends to strengthen which has led to speculations about the fidelity of the current climate models (Chung et al. 2019). On multidecadal and centennial timescales, it is *likely* that mean ITF decreases which is not associated with wind variability but due to reduction of net deep ocean upwelling in the tropical South Pacific (Sen Gupta et al. 2016; Feng et al. 2017; Feng et al. 2018). Due to a lack of long-term sustained ITF observations, their impacts on Indo-Pacific climate varibility, biogeochemisty, ecosystem as well as society are not fully understood.

Pacific SST cooling trends and strengthened the equatorial Pacific trade winds have been linked to anomalously warm tropical Indian and Atlantic oceans. The period following the mid-1990s saw a marked strengthening of both the easterly trade winds in the central equatorial Pacific (Figure 6.7) and the Walker circulation (L'Heureux et al. 2013; England et al. 2014). Both the magnitude and duration of this trend are large when compared with past variability reconstructed using atmosphere reanalyses. (The 1886–1905 extreme weakening trend is poorly constrained by observations and we note the disparity

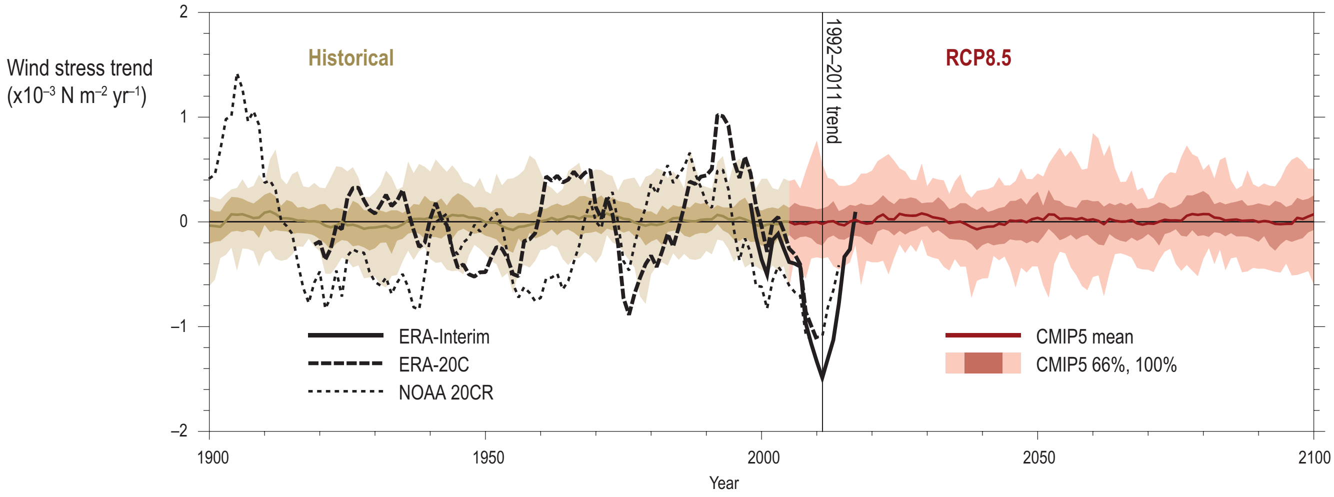

Figure 6.7 | Running twenty-year trends of zonal wind stress over the central Pacific (area-averaged over 8ºS–8ºN and 160ºE–150ºW) in Coupled Model Intercomparison Project Phase 5 (CMIP5) models and three reanalyses: European Centre for Medium-Range Weather Forecasts (ECMWF) Interim re-analysis, ERA-Interim (Dee et al. 2011), ECMWF 20th century reanalysis, ERA-20C (Poli et al. 2016), and the National Oceanic and Atmospheric Administration's (NOAA) 20th century reanalysis, NOAA 20CR v2c (Compo et al. 2011). The 66% and 100% ranges of all available CMIP5 historical simulations with Representative Concentration Pathway (RCP)8.5 extension are shown.

between reanalysis products going back in time.) Moreover, it is very unusual when model simulations are used as an estimate of internal climate variability (Figure 6.7; England et al. 2014; Kociuba and Power, 2015). The slowdown in global surface warming is dominated by the cooling in the Pacific SSTs, which is associated with a strengthening of the Pacific trade winds (Kosaka and Xie, 2013). This pattern leads to cooling over land and possibly to additional heat uptake by the ocean, although recent studies suggest that ocean heat uptake may even slow down during surface warming slowdown periods (Xie et al. 2016; von Känel et al. 2017). The intensification of the Pacific trade winds has been related to inter-ocean basin SST trends, with rapid warming in the Indian (see section 6.5.1.2) and Atlantic Oceans both hypothesised as drivers (Kucharski et al. 2011; Luo et al. 2012; McGregor et al. 2014; Zhang and Karnauskas, 2017). While the extreme event of strengthening trade winds are potentially a result of natural internal variability, a role of anthropogenic contribution has not been ruled out. Nevertheless, the CMIP5 models indicate no general change in trends into the future (Figure 6.7), giving more weight to natural internal variability as an explanation.

Among the number of potential causes of this decadal variability in surface global temperature, a prolonged negative phase of the Pacific Decadal Oscillation/Interdecadal Pacific Oscillation (PDO/IPO) was suggested as a contributor. Because of the magnitude and duration of this Pacific-centred variability (Figure 6.7), it is identified as an extreme decadal climate event. One line of research has explored the role of the warm tropical Atlantic decadal variability in forcing the trade wind trends and associated cooling Pacific SST trends (Kucharski et al. 2011; McGregor et al. 2014; Li et al. 2016b). It appears that climate models may misrepresent this link due to tropical Atlantic biases (Kajtar et al. 2018; McGregor et al. 2018) and thus potentially underestimate global mean temperature decadal variability. Nevertheless, there is no indication that such an underestimation of global temperature variability is evident in the models (Flato et al. 2013; Marotzke and Forster, 2015). The impact of modes of natural variability on global mean temperature decadal variability remains an active area of research.

In the Indian Ocean, water exits the Indonesian Seas mostly flowing westward along with the South Equatorial Current, and some supplying the Leeuwin Current. The South Equatorial Current feeds the heat and biogeochemical signatures from the Indian Ocean into the Agulhas Current, which transports it further into the Atlantic Ocean. Observations of Mozambique Channel inflow from 2003–2012 measured a mean transport of 16.7 x 10^6 m^3 s^{-1} with a maximum in austral winter, and IOD related interannual variability of 8.9 x 10^6 m^3 s^{-1} (Ridderinkhof et al. 2010). A multidecadal proxy, from three years of mooring data and satellite altimetry, suggests that the Agulhas Current has been broadening since the early 1990s due to an increase in eddy kinetic energy (Beal and Elipot, 2016). Numerical model experiments suggest an intensification of the Agulhas leakage since the 1960s, which has contributed to the warming in the upper 300 m of the tropical Atlantic Ocean (Lübbecke et al. 2015). Agulhas leakage is found to covary with the AMOC on decadal and multi-decadal timescales and has *likely* contributed to the AMOC slowdown (Biastoch et al. 2015; Kelly et al. 2016). Meanwhile, climate projections indicate that Agulhas leakage is *likely* to strengthen and may partially compensate the AMOC slowdown projected by coarse-resolution climate models (Loveday et al. 2015).

6.6.2 Impacts on Natural and Human Systems

Interannual to decadal variability of Indo-Pacific SST variability is *likely* to affect extreme hydroclimate in East Africa (Ummenhofer et al. 2018). The Pacific cooling pattern is often synonymous with predominance of La Niña events in 1998 and 2012 is linked to megadroughts in the USA (Baek et al. 2019). On decadal to multidecadal time scales, PDO/IPO and Atlantic variability may have impacts on megadroughts in North America (Coats et al. 2016; Diodato et al. 2019) and Australia (Vance et al. 2015) as well as Indian subcontinent (Bao et al. 2015; Joshi and Rai, 2015). It is *likely* that occurrence of megadroughts in North America and Australia increased (Kiem et al. 2016; Baek et al. 2019). PDO and North Pacific Gyre Oscillation may also influence the decadal variability of North Pacific nutrient, chlorophyll and zooplankton taxa (Di Lorenzo et al. 2013).

The Pacific cooling pattern may have significant impacts on terrestrial carbon uptake via teleconnections. The reduced ecosystem respiration due to the smaller warming over land has significantly accelerated the net biome productivity and therefore increased the terrestrial carbon sink (Ballantyne et al. 2017) and paused the growth rate of atmospheric CO_2 despite increasing anthropogenic carbon emissions (Keenan et al. 2016). During the 2000s, the global ocean carbon sink has also strengthened (Fay and McKinley, 2013; Landschützer et al. 2014; Majkut et al. 2014; Landschützer et al. 2015; Munro et al. 2015), reversing a trend of stagnant or declining carbon uptake during the 1990s. It has been suggested that the upper ocean overturning circulation has weakened during the 2000s thereby decreasing the outgassing of natural CO_2, especially in the Southern Ocean (Landschützer et al. 2015), and enhanced the global ocean CO_2 sink (DeVries et al. 2017). How this is connected to the global warming slowdown is currently unclear.

6.7 Risks of Abrupt Change in Ocean Circulation and Potential Consequences

6.7.1 Key Processes and Feedbacks, Observations, Detection and Attribution, Projections

6.7.1.1 Observational and Model Understanding of Atlantic Ocean Circulation Changes

Palaeo-reconstructions indicate that the North Atlantic is a region where rapid climatic variations can occur (IPCC, 2013). Deep waters formed in the northern North Atlantic induces a large-scale AMOC which transports large amounts of heat northward across the hemispheres, explaining part of the difference in temperature between the two hemispheres, as well as the northward location of the ITCZ (e.g., Buckley and Marshall, 2016). This circulation system is believed to be a key tipping point of the Earth's climate system (IPCC, 2013).

Considerable effort has been dedicated in the last decades to improve the observation system of the large-scale ocean circulation (e.g., Argo and its array of about 3,800 free-drifting profiling floats), including the AMOC through dedicated large-scale observing arrays (at 16ºN (Send et al. 2011) and 26ºN (McCarthy et al. 2015b), in the subpolar gyre (SPG) (Lozier et al. 2017), between Portugal and the tip of Greenland (Mercier et al. 2015), at 34.5ºS (Meinen et al. 2013), among others). The strength of the AMOC at 26ºN has been continuously estimated since 2004 with an annual mean estimate of 17 ± 1.9 x 10^6 m^3 s^{-1} over the 2004–2017 period (Smeed et al. 2018). The AMOC at 26ºN has been 2.7 x 10^6 m^3 s^{-1} weaker in 2008–2017 than in the first four years of measurement (Smeed et al. 2018). However, the record is not yet long enough to determine if there is a long-term decline of the AMOC. McCarthy et al. (2012) reported a 30% reduction in the AMOC in 2009–2010, followed by a weaker minimum a year later. Analysis of forced ocean models suggests such events may occur once every two or three decades (Blaker et al. 2015). At 34.5ºS, the mean AMOC is estimated as 14.7 ± 8.3 x 10^6 m^3 s^{-1} over the period 2009–2017 (Meinen et al. 2018) also with large interannual variability, while no trend has been identified at this latitude. Estimates based on ocean reanalyses show considerable diversity in their AMOC mean state, and its evolution over the last 50 years (Karspeck et al. 2017; Menary and Hermanson, 2018), because only very few deep ocean observations before the Argo era, starting around 2004, are available. During the Argo era, the reanalyses agree better with each other (Jackson et al. 2016).

During the last interglacial warm period, palaeo-data suggest that the AMOC may have been weaker (Govin et al. 2012) and also show proxy record evidences of instabilities (Galaasen et al. 2014). Based on an AMOC reconstruction using SST fingerprints, it has been suggested that the AMOC may have experienced around 3 ± 1 x 10^6 m^3 s^{-1} of weakening (about 15% decrease) since the mid-20th century (Caesar et al. 2018). Such a trend in AMOC was also suspected in a former study using Principal Component Analysis of SST (Dima and Lohmann, 2010). Palaeo-proxies also highlight that the historical era may exhibit an unprecedented low AMOC over the last 1,600 years (Sherwood et al. 2011; Rahmstorf et al. 2015; Thibodeau et al. 2018; Thornalley et al. 2018). Nevertheless, these proxy records are indirect measurements of the AMOC so that considerable uncertainty remains concerning these results. Moreover, the exact mechanisms to explain such a long-term weakening are not fully understood and some reconstructions show a weakening starting very early in the historical era, when the level of anthropogenic perturbation and warming was very low. Climate model simulations (Figure 6.8) do show a weakening over the historical era, but this weakening is mainly occurring over the recent decades. Climate projections exhibit a weakening of around 1.4 ± 1.4 x 10^6 m^3 s^{-1} for present day (2006–2015) minus pre-industrial (1850–1900), highlighting that anthropogenic warming may have already forced an AMOC weakening. Nevertheless, no proper detection and attribution of the on-going changes has been led so far due to still limited observational evidences. Thus, we conclude that there is *medium confidence* that the AMOC has weakened over the historical era but there is insufficient evidence to quantify a *likely* range of the magnitude of the change.

Examination of 14 models from the CMIP5 archive, which do not take into account the melting (either from runoff, basal melting or icebergs) from the GIS (cf. Section 6.7.1.2), led to the assessment that the AMOC is *very unlikely* to collapse in the 21st century in response to increasing GHG concentrations (IPCC, 2013). Nonetheless, the CMIP5 models agree that a weakening of the AMOC into the 21st century will lead to localised cooling (relative to the global mean) centred in the North Atlantic SPG (Menary and Wood, 2018), although the precise location as well as the extension of this cooling patch, notably towards Europe, remains uncertain (Sgubin et al. 2017; Menary and Wood, 2018).

Abrupt variations in SST or sea ice cover have been found in 19 out of the 40 models of the CMIP5 archive (Drijfhout et al. 2015). Large cooling trends, which can occur in a decade, are found in the subpolar North Atlantic in 9 out of 40 models. Results show that the heat transport in the AMOC plays a role in explaining such a rapid cooling, but other processes are also key for setting the rapid (decadal-scale) timeframe of SPG cooling, notably vertical heat transport in the ocean and interactions with sea ice and the atmosphere (Sgubin et al. 2017). Using the representation of stratification as an emergent constraint, rapid changes in subpolar convection and associated cooling are occurring in the 21st century in 5 of the 11 best models (Sgubin et al. 2017). The poor representation of ocean deep convection in most CMIP5 models has been confirmed in Heuze (2017), which can notably limit a key feedback mechanism related with warm summer in the North Atlantic and its impact on oceanic convection in winter (Oltmanns et al. 2018). Thus, there is *low confidence* in the projections of SPG fate. Increasing the horizontal resolution of the ocean in next generation climate models might be a way to increase confidence in ocean convection future changes.

The SPG dynamical system has been identified as a tipping element of the climate system (Mengel et al. 2012; Born et al. 2013). If this element reaches its tipping point, the SPG circulation can change very abruptly between different stable steady states, due to positive feedback between convective activity and salinity transport within the gyre (Born et al. 2016). It has been argued that a transition between two SPG stable states can explain the onset of the Little Ice Age that may have occurred around the 14–15th century (Lehner et al. 2013; Schleussner et al. 2015; Moreno-Chamarro et al. 2017) possibly triggered by large volcanic eruption (Schleussner and Feulner, 2013). Furthermore a few CMIP5 climate models also showed a rapid cooling in the SPG within the 1970s cooling events, as a nonlinear response to aerosols (Bellucci et al. 2017). The SPG therefore appears as a tipping element in the climate system, with a faster (decade) response than the AMOC (century), but with lower induced SST cooling. Thus, the SPG system can cross a threshold in climate projections when surface water in the subpolar becomes lighter due to increase in temperature and decrease in salinity related with changes in radiative forcing (Sgubin et al. 2017).

Evaluation of AMOC variations in the CMIP5 database has been further analysed in this report (Figure 6.8) using almost twice as many models as in AR5 (IPCC, 2013). The AR5 assessment of a *very unlikely* AMOC collapse has been confirmed, although one model (FGOALS-s2) does show such a collapse (e.g., decrease larger than 80% relative to present day) before the end of the century for RCP8.5 scenario (Figure 6.8). Now based on up to 27 model simulations, the decrease of the AMOC is assessed to be of -2.1 ± 2.6 x 10^6 m^3 s^{-1} (-11 ± 14%, *likely* range) in 2081–2100 relative to present day (2006–2015) for RCP2.6 scenario and -5.5 ± 2.7 x 10^6 m^3 s^{-1} (-32 ± 14%) for RCP8.5 scenario, in line with a process-based probabilistic assessment (Schleussner et al. 2014). Furthermore, the uncertainty in AMOC changes has been shown to be mainly related to the spread in model responses rather than scenarios (RCP4.5 and RCP8.5) or internal variability uncertainty (Reintges et al. 2017). This behaviour is very different from the uncertainty in global SST

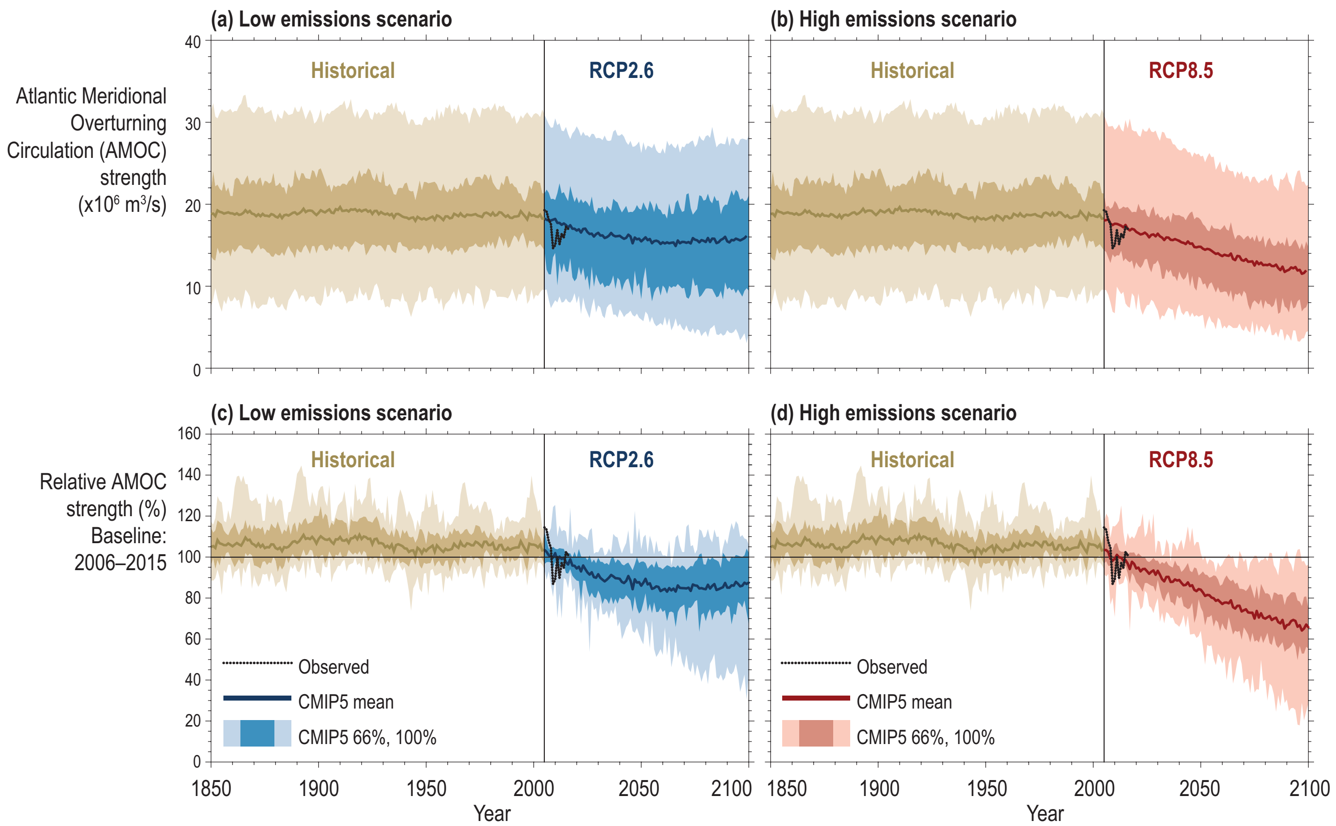

Figure 6.8 | Atlantic Meridional Overturning Circulation (AMOC) changes at 26°N as simulated by 27 models (only 14 were shown in the IPCC 5th Assessment Report (AR5); IPCC, 2013). The dotted line shows the observation-based estimate at 26°N (McCarthy et al. 2015b) and the thick grey/blue/red lines the multi-model ensemble mean. Values of AMOC maximum at 26°N (in units 10^6 m^3 s^{-1}) are shown in historical simulations (most of the time 1850–2005) followed for 2006–2100 by **a)** Representative Concentration Pathway (RCP)2.6 simulations and **b)** RCP8.5 simulations. In c) and d), the time series show the AMOC strength relative to the value during 2006–2015, a period over which observations are available. **c)** shows historical followed by RCP2.6 simulations and **d)** shows historical followed by RCP8.5 simulations. The 66% and 100% ranges of all-available CMIP5 simulations are shown in grey for historical, blue for RCP2.6 scenario and red for RCP8.5 scenario.

changes, which is mainly driven by emission scenario after a few decades (Frölicher et al. 2016). To explain the AMOC decline, a new mechanism has been proposed on top of the classical changes in heat and freshwater forcing (Gregory et al. 2016). A potential role for sea ice decrease has been highlighted (Sevellec et al. 2017), due to large heat uptake increase in the Arctic leading to a strong warming of the North Atlantic, increasing the vertical stability of the upper ocean, as already observed in the Greenland and Iceland seas (Moore et al. 2015). It has also been showed that convection sites may move northward in future projections, following the sea ice edge (Lique and Thomas, 2018).

6.7.1.2 Role of GIS Melting and their Freshwater Release Sources

Satellite data indicate accelerated mass loss from the GIS beginning around 1996, and freshwater contributions to the subpolar North Atlantic from Greenland, Canadian Arctic Archipelago glaciers and sea ice melt totalling around 60,000 $m^3 s^{-1}$ in 2013, a 50% increase since the mid-1990s (Yang et al. 2016b), in line with more recent estimates (Bamber et al. 2018). This increase in GIS melting is unprecedented over the last 350 years (Trusel et al. 2018). Since the mid-1990s, there has been about a 50% decrease in the thickness of the dense water mass formed in the Labrador Sea, suggesting a possible relationship between enhanced freshwater fluxes and suppressed formation of North Atlantic Deep Water (Yang et al. 2016b). This hypothesis has been further supported by high-resolution ocean-only simulations showing that GIS melting may have affected the Labrador Sea convection since 2010, which may imply an emerging on-going impact of this melting on the SPG but a still non-detectable impact on the AMOC (Boning et al. 2016). Thus, while some studies argue that this melting may have affected the evolution of the AMOC over the 20th century (Rahmstorf et al. 2015; Yang et al. 2016b), considerable variability and limitation in ocean models restrain the full validation of this hypothesis, which remains model dependent (Proshutinsky et al. 2015; Dukhovskoy et al. 2016). Furthermore, some deep convection events resumed since 2014 (Yashayaev and Loder, 2017).

The impact of GIS melting is neglected in AR5 projections (Swingedouw et al. 2013) but has been considered in a recent multi-model study (Bakker et al. 2016; Figure 6.9). The decrease of the AMOC in projections including this melting term is depicted in Figure 6.9. GIS melting estimates added in those simulations were based on the Lenaerts et al. (2015) approach, using a regional atmosphere model to estimate GIS mass balance. Results from eight climate models and an extrapolation by an emulator calibrated on these models showed that GIS melting has an impact on the AMOC, potentially adding up to around 5–10% more AMOC weakening in

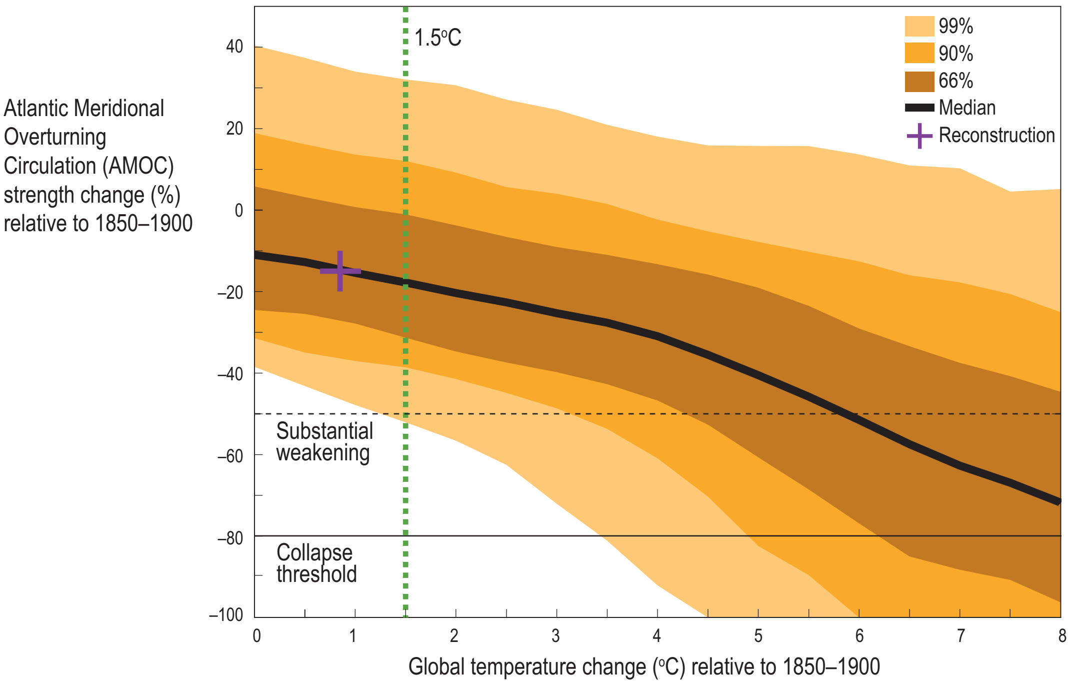

Figure 6.9 | The changes in the Atlantic Meridional Overturning Circulation (AMOC) strength as a function of transient changes in global mean temperature for projections from RCP4.5 and RCP8.5 scenario. This probabilistic assessment of annual mean AMOC strength changes (%) at 26ºN (below 500 m and relative to 1850–1900) as a function of global temperature change (degrees Celsius; relative to 1850–1900) results from 10,000 RCP4.5 and 10,000 RCP8.5 experiments over the period 2006–2300, which are derived from an AMOC emulator calibrated with simulations from eight climate models including the Greenland Ice Sheet (GIS) melting (Bakker et al. 2016). The annual mean AMOC strength changes are taken from transient simulations and are therefore not equilibrium values *per se*. Moreover, it should be stressed that the results stem from future runs, not past or historical runs. Thus, due to internal variability both in the global mean temperature and AMOC in this transient simulation, large weakening can be found even at 0ºC global warming. The ranges (66%, 90% and 99%) correspond to the amount of simulations that are within each envelope. The thick black line corresponds to the ensemble mean, while the different colours stand for different probability quantiles. The horizontal black thick line corresponds to the value of 80% of AMOC decrease, which can be seen as an almost total collapse of the AMOC. The horizontal black dashed thick line corresponds to a reduction of 50% of the AMOC, which can be considered as a substantial weakening. The vertical dashed green line stands for the 1.5ºC of global warming threshold (relative to 1850–1900). The violet cross stands for the observation-based reduction estimate from Caesar et al. (2018). The size of the cross represents the uncertainty in this estimate.

2100 under RCP8.5. Based on Figure 6.8 and 6.9, the risk of collapse before the end of the century is *very unlikely*, although biases in present-day climate models only provide *medium confidence* in this assessment. By 2290–2300, Bakker et al. (2016; Figure 6.9) estimated at 44% the likelihood of an AMOC collapse in RCP8.5 scenario, while the AMOC weakening stabilises in RCP4.5 (37% reduction, (15–65%) *very likely* range). This result suggests that an AMOC collapse can be avoided in the long term by mitigation.

Concerning the question of the reversibility of the AMOC, a few ramp-up/ramp-down simulations have been performed to evaluate it for transient time scales (a few centuries, while millennia will be necessary for a full steady state). Results usually show a reversibility of the AMOC (Jackson et al. 2014; Sgubin et al. 2015) although the timing and amplitude is highly model dependent (Palter et al. 2018). A hysteresis behaviour of the AMOC in response to freshwater release has been found in a few climate models (Hawkins et al. 2011; Jackson et al. 2017) even at the eddy resolving resolution (Mecking et al. 2016; Jackson and Wood, 2018). This is in line with the possibility of tipping point in the AMOC system. The biases of present-day models in representing the transport at 30ºS (Deshayes et al. 2013; Liu et al. 2017a; Mecking et al. 2017) or the salinity in the tropical era (Liu et al. 2014b) may considerably affect the sensitivity of the models to freshwater release, but more on the multi-centennial time scale.

Regarding the near-term changes of the AMOC, decadal prediction systems are now in place. They indicate a clear impact of the AMOC on the climate predictability horizon (Robson et al. 2012; Persechino et al. 2013; Robson et al. 2013; Wouters et al. 2013; Msadek et al. 2014; Robson et al. 2018), and a possible weakening of the AMOC in the coming decade (Smith et al. 2013; Hermanson et al. 2014; Yeager et al. 2015; Robson et al. 2016), although not true in all decadal prediction systems (Yeager et al. 2018). All these prediction systems do not account for future melting of the GIS yet.

6.7.2 Impacts on Climate, Natural and Human Systems

Even though the AMOC is *very unlikely* to collapse over the 21st century, its weakening may be substantial, which may therefore induce strong and large-scale climatic impacts with potential far-reaching impacts on natural and human systems (e.g., Good et al. 2018). Furthermore, the SPG subsystem has been shown to potentially shift, in the future, into a cold state over a decadal time scale, with significant climatic implications for the North Atlantic bordering regions (Sgubin et al. 2017). There have been far more studies analysing impacts on climate of an AMOC weakening than SPG collapse. We will thus in the following mainly depict impacts of an AMOC substantial weakening.

The AR5 report concludes that based on palaeoclimate data, large changes in the Atlantic Ocean circulation can cause worldwide climatic impacts (Masson-Delmotte et al. 2013), with notably, for an AMOC weakening, a cooling of the North Atlantic, a warming of the South Atlantic, less evaporation and therefore precipitation over the North Atlantic, and a shift of the ITCZ. Impacts of AMOC or SPG changes and their teleconnections in the atmosphere and ocean are supported by a large amount of palaeo-evidence (Lynch-Stieglitz, 2017). Such impacts and teleconnections have been further evaluated over the last few years both using new palaeo-data and higher resolution models. Furthermore, multi-decadal variations in SST observed over the last century, the so-called Atlantic Multidecadal Variability (AMV) or Atlantic Multidecadal Oscillation (AMO), also provide observational evidence of potential impacts of changes in ocean circulation. Nevertheless, due to a lack of long-term direct measurements of the Atlantic Ocean circulation, the exact link between SST and circulation remains controversial (Clement et al. 2015; Zhang, 2017).

The different potential impacts of large changes in the Atlantic Ocean circulation are summarised in Figure 6.10. Based on variability analysis, it has been shown that a decrease in the AMOC strength has impacts on storm track position and intensity in the North Atlantic (Gastineau et al. 2016), with a potential increase in the number of winter storms hitting Europe (Woollings et al. 2012; Jackson et al. 2015), although some uncertainty remains with respect to the models considered (Peings et al. 2016). The influence on the Arctic sea ice cover has also been evidenced at the decadal scale, with a lower AMOC limiting the retreat of Arctic sea ice (Yeager et al. 2015; Delworth and Zeng, 2016). The climatic impacts could be substantial over Europe (Jackson et al. 2015), where an AMOC weakening can lead to high pressure over the British Isles in summer (Haarsma et al. 2015), reminiscent of a negative summer NAO, inducing an increase in precipitation in Northern Europe and a decrease in Southern Europe. In winter, the response of atmospheric circulation may help to reduce the cooling signature over Europe (Yamamoto and Palter, 2016), notably through an enhancement of warming maritime effect due to a stronger storm track (Jackson et al. 2015), driving more powerful storms in the North Atlantic (Hansen et al. 2016). The observed extreme low AMOC in 2009–2010, which was followed by a reduction in ocean heat content to the north (Cunningham et al. 2013), has been possibly implicated in cold European weather events in winter 2009–2010 and December 2010 (Buchan et al. 2014) although a robust attribution is missing. In summer, cold anomalies in the SPG, like the one occurring during the so-called cold blob in 2015 (Josey et al. 2018), have been suspected to potentially enhance the probability of heatwaves over Europe in summer (Duchez et al. 2016). Nevertheless, considerable uncertainties remain with regard to this aspect due to the lack of historical observations before 2004 and due to poor model resolution of small-scale processes related to frontal dynamics around the Gulf Stream region (Vanniere et al. 2017). In addition, oceanic changes in the Gulf Stream region may occur in line with AMOC weakening (Saba et al. 2016) with potential rapid warming due to a northward shift of the Gulf Stream. However, these changes are largely underestimated in coarse resolution models (Saba et al. 2016) . In North America, a negative phase of the AMV, reminiscent of a weakening of the AMOC, lowers agricultural production in a few Mexican coastal states (Azuz-Adeath et al. 2019).

Changes in ocean circulation can also strongly impact sea level in the regions bordering the North Atlantic (McCarthy et al. 2015a; Palter et al. 2018). A collapse of the AMOC or of the SPG could induce substantial increase of sea level up to a few tens

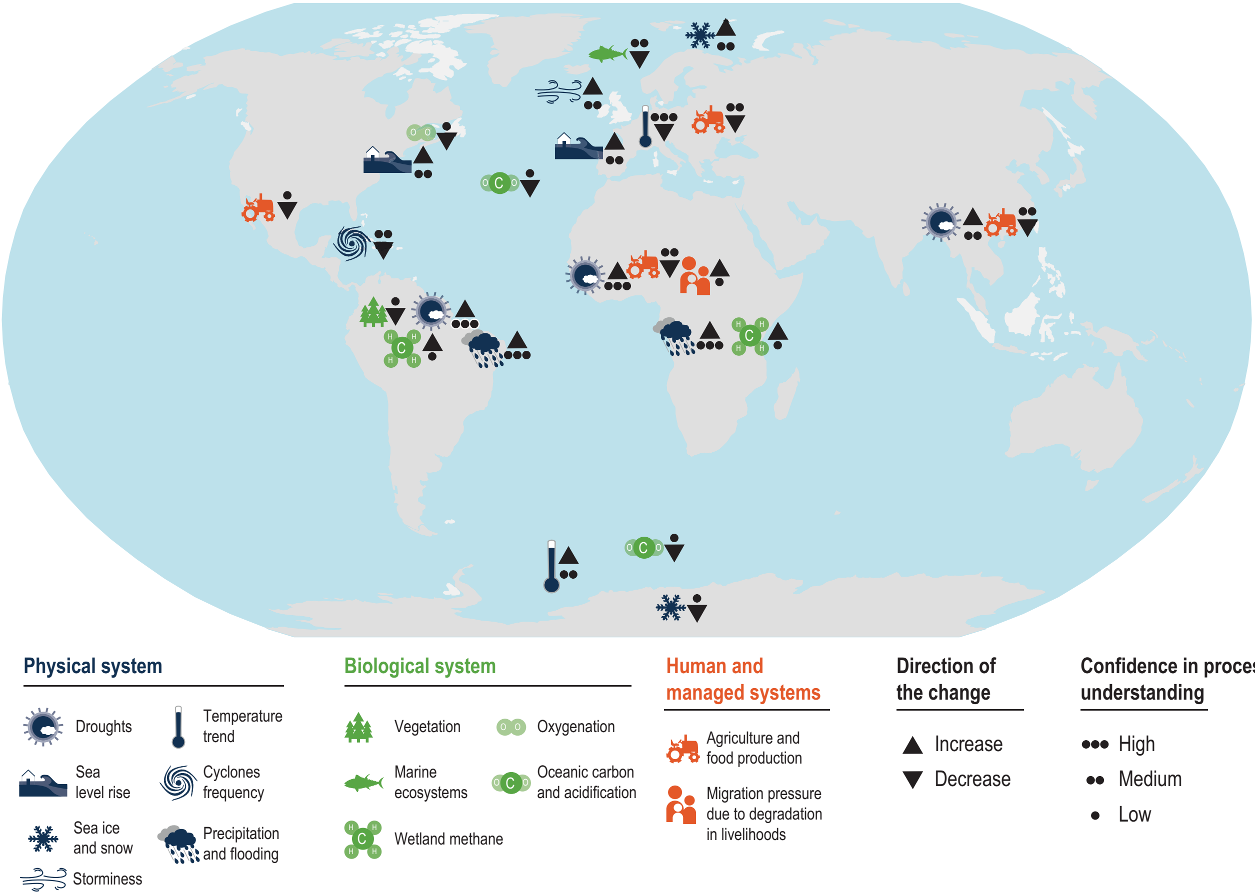

Figure 6.10 | Infographic on teleconnections and impacts due to Atlantic Meridional Overturning Circulation (AMOC) collapse or substantial weakening. Changes in circulation have multiple impacts around the Atlantic Basin, but also include remote impacts in Asia and Antarctica. Reductions in AMOC lead to an excess of heat in the South Atlantic, leading to increased flooding, methane emissions and drought, and a concomitant negative impact on food production and human systems. In the North Atlantic region hurricane frequency is decreased on the western side of the basin, but storminess increases in the east. Marine and terrestrial ecosystems, including food production, are impacted while sea level rise (SLR) is seen on both sides of the Atlantic. The arrows indicate the direction of the change associated with each icon and is put on its right. An assessment of the confidence level in the understanding of the processes at play is indicated below each arrow.

of centimetres along the western boundary of the North Atlantic (Ezer et al. 2013; Little et al. 2017; cf. Chapter 5). For instance, such a link may explain 30% of the extreme observed SLR event (a short-lived increase of 12 mm during 2 years) in northeast America in 2009–2010 (Ezer, 2015; Goddard et al. 2015). This illustrates that monitoring changes in AMOC may have practical implications for coastal protection.

The AMOC teleconnections are widespread and notably strongly affect the tropical area, as evidenced in palaeo-data for the Sahel region (Collins et al. 2017; Mulitza et al. 2017) and in model simulations (Jackson et al. 2015; Delworth and Zeng, 2016). These teleconnections may affect vulnerable populations. For instance, Defrance et al. (2017) found that a substantial decrease in the AMOC, at the very upper end of potential changes, may strongly diminish precipitation in the Sahelian region, decreasing the millet and sorghum emblematic crop production, which may impact subsistence of tens of millions of people, increasing their potential for migration. Smaller amplitude variations in Sahelian rainfall, driven by North Atlantic SST, has been found to be predictable up to a decade ahead (Gaetani and Mohino, 2013; Mohino et al. 2016; Sheen et al. 2017), potentially providing mitigation and adaptation opportunities. The number of tropical storms in the North Atlantic has been found to be very sensitive to the AMOC (Delworth and Zeng, 2016; Yan et al. 2017) as well as to the SPG (Hermanson et al. 2014) variations, so that a large weakening of the AMOC or cooling of the SPG may decrease the number of Atlantic tropical storms. The Asian monsoon may also potentially weaken in the case of large changes in the AMOC (Marzin et al. 2013; Jackson et al. 2015; Zhou et al. 2016; Monerie et al. 2019) implying substantial adverse impacts on populations. The interactions of the Atlantic basin with the Pacific has also been largely discussed over the last few years, with the supposed influence of a cool North Atlantic inducing a warm tropical Pacific (McGregor et al. 2014; Chafik et al. 2016; Li et al. 2016b), although not found in all models (Swingedouw et al. 2017), which may induce stronger amplitudes of El Niño (Dekker et al. 2018).

The AMOC plays an important function in transporting excess heat and anthropogenic carbon from the surface to the deep ocean (Kostov et al. 2014; Romanou et al. 2017), and therefore in setting the pace of global warming (Marshall et al. 2014). A large potential decline in the AMOC strength reduces global surface warming. This is due to changes in the location of ocean heat uptake and associated expansion of the cryosphere around the North Atlantic, which increases surface albedo (Rugenstein et al. 2013; Winton et al. 2013), as well as cloud cover variations and modifications in water vapour content (Trossman et al. 2016). As the uptake of excess heat occurs preferentially in regions with delayed warming (Winton et al. 2013; Frölicher et al. 2015; Armour et al. 2016), a potential large reduction of the AMOC may shift the uptake of excess heat from the low to the high latitudes (Rugenstein et al. 2013; Winton et al. 2013), where the atmosphere is more sensitive to external forcing (Winton et al. 2010; Rose et al. 2014; Rose and Rayborn, 2016; Rugenstein et al. 2016). A decrease in AMOC may also decrease the subduction of anthropogenic carbon to deeper waters (Zickfeld et al. 2008; Winton et al. 2013; Randerson et al. 2015; Rhein et al. 2017). A potential impact of methane emissions has also been highlighted for past Heinrich events during which massive icebergs discharge in the North Atlantic may have led to large AMOC disruptions. Large increases (>100 ppb) in methane production have been associated with these events (Rhodes et al. 2015) potentially due to increased wetland production in the SH, related to teleconnections of the North Atlantic with tropical area (Ringeval et al. 2013; Zurcher et al. 2013). All these different effects indicate a potentially positive feedback of the AMOC on the carbon cycle (Parsons et al. 2014), although other elements from the terrestrial biosphere may limit its strength or even reverse its sign (Bozbiyik et al. 2011).

Changes in Atlantic Ocean circulation can also strongly impact marine life and can be seen at all levels of different ecosystems. For instance, changes in the abundance and distribution of species in response to circulation changes in the SPG have been documented amongst plankton (Hátún et al. 2009), fish (Payne et al. 2012; Miesner and Payne, 2018), seabirds (Descamps et al. 2013) and top predators such as tuna, billfish and pilot whales (Hátún et al. 2009; MacKenzie et al. 2014). Nutrient concentrations in the northeast Atlantic have also been shown to be limited by the recent weakening of the SPG, with concomitant ecosystem impacts (Johnson et al. 2013; Hátún et al. 2016). The influence of SPG circulation also extends to ecosystems beyond from the immediate area, and has a clear impact on the productivity of cod (*Gadus morhua*) in the Barents Sea, for example (Årthun et al. 2017; Årthun et al. 2018). On a broader scale, changes in the AMOC are an important driver of AMV, which has also been linked to substantial changes in marine ecosystems on both sides of the North Atlantic (Alheit et al. 2014; Nye et al. 2014). Recent AMOC weakening is also suspected to explain large marine deoxygenation in the northwest coastal Atlantic (Claret et al. 2018). In addition, a recent study using a marine productivity proxy from Greenland ice cores suggest that net primary productivity has decreased by 10 ± 7% in the subarctic Atlantic over the past two centuries possibly related to changes in AMOC (Osman et al. 2019). Finally, a model study investigated the impact of mitigation by reversing the forcing from a RCP8.5 scenario from 2100 and found that global marine net productivity may recover very rapidly and even overshoot contemporary values at the end of the reversal, highlighting the potential benefit of mitigation (John et al. 2015).

Following all these potential impacts, it has been suggested that a collapse of the AMOC may have the potential to induce a cascade of abrupt events, related to the crossing of thresholds from different tipping points, itself potentially driven by GIS rapid melting. For example, a collapse of the AMOC may induce causal interactions like changes in ENSO characteristics (Rocha et al. 2018), dieback of the Amazon rainforest and shrinking of the WAIS due to seesaw effect, ITCZ southern migration and large warming of the Southern Ocean (Cai et al. 2016). However, such a worst case scenario remains very poorly constrained quantitatively due to the large uncertainty in GIS and AMOC response to global warming.

The potential impacts of such rapid changes in ocean circulation on agriculture, economy and human health remain poorly evaluated up to now with very few studies on the topic (Kopits et al. 2014). The available impact literature on AMOC weakening has focussed on impacts from temperature change only (reduced warming), globally leading to economic benefits (e.g., Anthoff et al. 2016), and local losses can amount to a few percent of gross domestic product (GDP), however under a complete shutdown (Link and Tol, 2011). Declines in Barents Sea fish species could lead to economic losses (Link and Tol, 2009), but more comprehensive economic studies are lacking.

6.7.3 Risk Management and Adaptation

The numerous potential impacts of AMOC weakening (see Section 6.7.2) require adaptation responses. A specific adaptation action is a monitoring and early warning system using observation and prediction systems, which can help to respond in time to effects of an AMOC decline. Although it is difficult to warn very early for large changes in AMOC to come, notably due to large natural decadal variability of the AMOC (Boulton et al. 2014), the observation arrays that are in place may allow the development of such an early warning system. Nevertheless, the prospects for its operational use for early warnings have not yet been fully developed. In this respect, developing early warning systems that do not depend on statistical timeseries analysis of long observational record might be seen as an important research goal in the future.

Decadal prediction systems can help fill this gap. Skilful prediction of AMOC variation has been demonstrated on the multi-annual scale (Matei et al. 2012) and retrospective prediction experiments have demonstrated that the large changes in the SPG seen in the mid-1990s could have been foreseen several years in advance (Wouters et al. 2013; Msadek et al. 2014). The World Climate Research Programme's grand challenge of launching decadal predictions every year (Kushnir et al. 2019) is an important step towards anticipating rapid changes in the near term and can drive decadal-scale climate services. For example, a few studies have already shown that small variations anticipated by decadal predictions (e.g., Sheen et al. 2017) can be useful for the development of climate services, notably for agriculture in south and east Africa (Nyamwanza et al. 2017). Decadal predictions also match the decision making time horizons of many users of the ocean (Tommasi et al. 2017b) and are expected to play an increasingly important role in this sector in the future (Payne et al. 2017).

6

6.8 Compound Events and Cascading Impacts

6.8.1 Concepts

Compound events refer to events that are characterised by multiple failures that can amplify overall risk and/or cause cascading impacts (Helbing, 2013; Gallina et al. 2016; Figure 6.1). These impacts may be triggered by multiple hazards that occur coincidently or sequentially and can lead to substantial disruption of natural or human systems (Leonard et al. 2014; Oppenheimer et al. 2014; Gallina et al. 2016; Zscheischler et al. 2018). These concepts are illustrated in a series of recent case studies that show how compound events interact with multiple elements of the ecosystem and society to create compound risk and cascading impacts (Box 6.1).

Compound events and cascading impacts are examples of deep uncertainty because data deficiency often prevents the assessment of probabilities and consequences of the risks from compound events. Furthermore, climate drivers that contribute to compound events could cross tipping points in the future (e.g., Cai et al. 2016; Cross-Chapter Box 4 in Chapter 1). Concepts and methods for addressing compound events and cascading impacts have a solid foundation in disaster risk reduction frameworks (Scolobig, 2017) where they may be assessed with scenarios, risk mapping, and participatory governance (Marzocchi et al. 2012; Komendantova et al. 2014). However, these approaches have tended to not consider the effects of climate change, rather considering hazards and vulnerability as stationary entities (Gallina et al. 2016). Trends in geophysical and meteorological extreme events and their interaction with more complex social, economic and environmental vulnerabilities overwhelm existing governance and institutional capacities (Shimizu and Clark, 2015) because of the aggregated cascading impacts.

6.8.2 Multiple Hazards

Understanding regions where changes in the climate system could increase the likelihood or severity of multiple hazards is relevant to understanding compound events (Figure 6.1). Several recent studies have highlighted coastal regions that are becoming more susceptible to multiple hazards from changes in regional climate. Warming and poleward expansion of the warm western boundary current regions (WBCs; Yang et al. 2016a) together with intensified cyclogenesis in these WBC regions; the Gulf Stream (Booth et al. 2012), the Kuroshio (Hirata et al. 2016) and the East Australian Current (EAC; Pepler et al. 2016a) can increase the likelihood of multiple hazards. These include increased rates of SLR (Brunnabend et al. 2017; Zhang et al. 2017b) together with increases in severe rainfall, storm surges and associated flooding (Thompson et al. 2013; Oey and Chou, 2016; Pepler et al. 2016a). WBCs have undergone an intensification and poleward expansion in all but the Gulf Stream where the weakening of the AMOC cancelled this effect (Seager and Simpson, 2016; Yang et al. 2016a).

Acknowledging the dual role of regional SLR and TCs frequency and intensity changes for future flood risk, Little et al. (2015) developed a flood index that takes account of local projected SLR along with TC frequency and intensity changes in a CMIP5 multi-model ensemble. They find that relative to 1986–2005, the Flood Index is 4–75 times higher by 2080–2099 for RCP2.6 (10–90th percentile range) and 35–350 times higher for RCP8.5. In the vicinity of the East Australian Current, Pepler et al. (2016b) found warmer SSTs boost the intensification of weak to moderate ETC's. Neglecting the compounding effects of flood and extreme sea level drivers can cause significant underestimation of flood risk and projected failure probability (Wahl et al. 2016; Moftakhari et al. 2017).

Over the last decade, several efforts have been made to address long-term shoreline change driven by the cascading impact of SLR, waves and MSL. Ranasinghe et al. (2012) presented the Probabilistic Coastline Recession model, which provides probabilistic estimates of coastline recession in response to both storms and SLR in the 21st century. Dune recession is estimated for each storm considering the recovery between storms, which is obtained empirically. More recently, Toimil et al. (2017) developed a methodology to address shoreline change over this century due to the action of waves, storm surges, astronomical tides in combination with SLR. The methodology considers the generation of thousands of multi-variate hourly time series of waves and storm surges to reconstruct future shoreline evolution probabilistically, which enables estimates of extreme recessions and long-term coastline change to be obtained. The model proposed by Vitousek et al. (2017) integrates longshore and cross-shore transport induced by GCM-projected waves and SLR, which allows it to be applied to both long and pocket sandy beaches. The analysis provides only one instance of what coastline change over the 21st century may be.

To summarise, new studies highlight regions such as coasts including those adjacent to WBCs, that are experiencing larger changes to multiple phenomena simultaneously such as SLR and cyclone intensity linked to higher SST increases (*medium confidence*), which increases the likelihood of extremes from multiple hazards occurring (*medium confidence*). Failing to account for the multiple factors responsible for extreme events will lead to an underestimation of the probabilities of occurrence (*high confidence*).

6.8.3 Cascading Impacts on Ecosystems

Damage and loss of ecosystems (mangrove, coral reefs, polar deserts, wetlands and salt marshes); or regime shifts in ecosystem communities lead to reduced resilience of all the ecosystems and possible flow-on effects to human systems. For example, recent studies showed that living corals and reef structures have experienced significant losses from human-related drivers such as coastal development; sand and coral mining; overfishing, acidification, and climate-related storms and bleaching events (Smith, 2011; Nielsen et al. 2012; Hilmi et al. 2013; Graham et al. 2015; Lenoir and Svenning, 2015; Hughes et al. 2017b). As a consequence, reef flattening is taking place globally due the loss of corals and from the bio-erosion and dissolution of the underlying reef carbonate structures (Alvarez-Filip et al. 2009). Reef mortality and flattening due to non-climate and climate-related drivers trigger cascading impacts and risks due to the loss of the protection services provided to coastal areas. High emission scenarios are expected to lead to almost the complete loss of coral cover by 2100, although policies aiming to lower the combined aerosol-radiation interaction and aerosol-cloud interaction (e.g., IPCC RCP 6.0) may partially limit the impacts on coral reefs and the associated habitat loss, thereby preserving an estimated 14 to 20 billion USD in consumer surplus 2100 (2014 USD, 3% discount; Speers et al. 2016). Moreover, projected SLR will increase flooding risks, and these risks will be even greater if reefs that now help protect coasts from waves are lost due to bleaching-induced mortality.

6.8.4 Cascading Impacts on Social Systems

Impacts of compound events also have significant multi-effects in the societal system. Cascading impacts are particularly driven by the loss or (temporary) disruption of critical infrastructure (Pescaroli and Alexander, 2018), such as communications, transport, and power supply, on housing, dams and flood protection; as well as health provision. Repeated extreme and compound events are leading to critical transitions in social systems (Kopp et al. 2016) which may cause the disruption of (local) communities, creating cascading impacts consisting of short-term impacts as well as long-lasting economic effects, and in some cases migration. When the responses of the economic sector to short term weather variations are applied to long term-climate projections, risks associated with climate change on different sectors are projected to result in an average 1.2% of decrease of US GDP per degree Celsius of warming. Furthermore, broad geographical discrepancies generate a large transfer of value northwards and westwards with the expected consequence of increased economic inequality (Hsiang et al. 2017). The severity and intensity of the cascading impacts also depend on the affected societies' vulnerability and resilience. For example, the intensity and influence of compound events are dependent on the size and scale of the affected society and the percentage of economy or GDP impacted (Handmer et al. 2012 in IPCC SREX). Smaller countries and especially small islands face the challenge of being unable to 'hedge' the risk through geographical redistribution (see Cross-Chapter Box 9).

Impacts from the natural system can descend into a cascade of disasters. For example, in 2005, Hurricane Katrina led to heavy flooding in the coastal area, dike breaches, emergency response failures, chaos in evacuation (traffic jams) and social disruption. Flooding in Thailand in 2011 led to the closure of many factories which not only impacted on the country's economy but impaired the global automobile and electronic industry (Kreibich et al. 2014). Female-owned establishments are more challenged with failures than businesses owned by men due to less experience, shorter duration and smaller size of businesses (Haynes et al. 2011; Marshall et al. 2015). The impact of compound events on ecosystems can also, in the long run, have devastating impacts on societal systems, for example, impacts from tropical storms can lead to coral degradation, which leads to increased wave impact and subsequent accelerated coastal erosion and impacts on fishing resources. This subsequently can have an impact on local economies, potentially leading to social disruption and migration (Saha, 2017). Impacts on marine ecosystems and habitats will also affect subsistence and commercial fisheries and, as a result, food security (Barrow et al. 2018). Climate-induced community relocations in Alaska stem from repeated extreme weather events coupled with climate change-induced coastal erosion and these impact the habitability of the whole community (Bronen, 2011; Durrer and Adams, 2011; Marino, 2011; Marino, 2012; Bronen and Chapin, 2013; see also Cross-Chapter Boxes 2 and 5 in Chapter 1).

6.8.5 Risk Management and Adaptation, Sustainable and Resilient Pathways

The management of compound events and cascading impacts in the context of governance poses challenges, partly because it is place dependent and heavily influenced by local parameters such as hazard experience and cultural values. Moreover, in some cases, people perceive that their community or country is less affected than others, leading to a 'spatial optimism bias' that delays or reduces the scope of actions (Nunn et al. 2016). In other cases it is unclear who will take responsibility when compound events and cascading impacts occur (Scolobig, 2017), although for some compound risks (e.g., na-tech disasters – when natural hazards trigger technological disasters), the private sector cooperates with governments to manage and respond to risks (Krausmann et al. 2017). Considerable variations exist among and inside countries. The level of engagement depends on the process of cascading impacts and the role of governance arrangement at the country level (Lawrence et al. 2018), countries' capacity to develop integrated risk and disaster frameworks and regulations, viable multi-stakeholder and public-private partnership in the case of multiple technological and natural hazards (Gerkensmeier and Ratter, 2018), the initiatives of local governments to exercise compound risk operations, and experience in interagency cooperation (Scolobig, 2017). The importance of local knowledge and traditional practices in disaster risk prevention and reduction is widely recognised (Hiwasaki et al. 2014; Hilhorst et al. 2015; Audefroy and Sánchez, 2017) (*high confidence*). The need to strengthen DRM is evident and can be improved and communicated effectively by integrating local knowledge such as Inuit's indigenous knowledge and local knowledge in Alaska (Pearce et al. 2015; Cross-Chapter Box 3 in Chapter 1) since it is easier for communities to accept than pure science-based DRM (Ikeda et al. 2016).

Despite difficulties of governance and decision making, many researchers and policy makers have recognised the need to study combined climatic and other hazards and their impacts. Several methods are now being employed to assess climatic hazards and compound events simultaneously, and also in combination (Klerk et al. 2015; van den Hurk et al. 2015; Wahl et al. 2015; Zscheischler and Seneviratne, 2017; Wu et al. 2018; Zscheischler et al. 2018). Policy makers can also begin to plan for disaster risk reduction and adaptation, based on these analyses of compound events and risks. Addressing limitations in understanding the compound hazards, as well as adequate mechanisms of the cascading impacts is needed. Finally, there are limits to resources to study these complex interactions in sufficient detail, as well as limits to data and information on past events that would allow the simulation of these effects, including economic impacts.

6.8.6 Global Impact of Tipping Points

A small number of studies (Lontzek et al. 2015; Cai et al. 2016; Lemoine and Traeger, 2016) use different versions of the Dynamic Integrated Climate-Economy assessment model (Nordhaus, 1992; Nordhaus, 2017) to assess the impact of diverse sets of tipping points and causal interactions between them on the socially optimal reduction of gas emissions and the present social cost of carbon, representing the economic cost caused by an additional ton of CO_2 emissions or its equivalent.

Cai et al. (2016) consider five interacting, stochastic, potential climate tipping points: reorganisation of the AMOC; disintegration of the GIS; collapse of the WAIS; dieback of the Amazon Rain Forest; and shift to a more persistent El Niño regime. The deep uncertainties associated with the likelihood of each of these tipping points and the dependence of them on the state of the others is addressed through expert elicitation. There *is limited evidence*, but *high agreement* that present costs of carbon are clearly underestimated. Double (Lemoine and Traeger, 2016), triple (Ceronsky et al. 2011), to eightfold (Cai et al. 2016) increase of the carbon price are suggested, depending on the working hypothesis. Cai et al. (2016) indicate that with the prospect of multiple interacting tipping points, the present social cost of carbon increases from 15 to 116 USD per tonne of CO_2, and conclude that stringent efforts are needed to reduce CO_2 emission if these impacts are to be avoided.

Box 6.1 | Multiple Hazards, Compound Risk and Cascading Impacts

The following case studies illustrate that anthropogenic climate change including ocean changes is increasingly having a discernible influence on elements of the climate system by exacerbating extreme events and causing multiple hazards, often with compound or sequential characteristics. In turn these elements are interacting with vulnerability and exposure to trigger compound events and cascading impacts.

Case Study 1: Tasmania's Summer of 2015–2016

Tasmania in southeast Australia experienced multiple extreme climate events in 2015–2016, driven by the combined effects of natural modes of climate variability and anthropogenic climate change, with impacts on the energy sector, fisheries and emergency services. The driest warm season on record (October to April), together with the warmest summer on record, brought agricultural and hydrological droughts to Tasmania and preconditioned the sensitive highland environment for major fires during the summer. Thousands of lightning strikes during the first two months of the year led to more than 165 separate vegetation fires, which burned more than 120,000 hectares including highland zones and the World Heritage Area and incurred costs to the state of more than 50 million AUD (Press, 2016).

In late January an intense cutoff low-pressure system brought heavy rainfall and floods, so that emergency services were simultaneously dealing with highland fires and floods in the east and north. The floods were followed by an extended wet period for Tasmania, with the wettest wet season (April to November) on record in 2016. Meanwhile, an intense marine heatwave (MHW) off the east coast persisted for 251 days from spring 2015 through to autumn 2016 (Oliver et al. 2017).

The driest October on record was influenced by both the El Niño and anthropogenic forcing (Karoly et al. 2016). Warmer sea surface temperatures (SSTs) due to anthropogenic warming may have increased the intensity of rainfall during the floods in January (e.g., Pepler et al. 2016a). The intensity and the duration of the MHW was unprecedented and both aspects had a clear human signature (Oliver et al. 2017).

Tasmania primarily relies on hydro-electric power generation and the trading of power over an undersea cable to mainland Australia, 'Basslink', for its energy needs. Lake levels in hydro-electric dams were at relatively low levels in early spring 2015, and the extended dry period led to further reductions and significantly reduced capacity to generate power (Hydro Tasmania, 2016). An unanticipated failure of the Basslink cable subsequently necessitated the use of emergency diesel generators (Hydro Tasmania, 2016).

The compound events caused many impacts on natural systems, agriculture, infrastructure and communities. Additional emergency services from outside the state were needed to deal with the fires. The MHW caused disease outbreaks in farmed shellfish, mortality in wild shellfish and species found further south than previously recorded. The energy sector experienced a severe cascade of impacts due to climate stressors and system inter-dependencies. The combination of drought, fires, floods and MHW reduced output from the agriculture, forestry, fishing and energy sectors and reduced the State of Tasmania gross state product (GSP) to 1.3%, well below the anticipated growth of 2.5%. To address the energy shortages, Tasmania's four largest industrial energy users, responsible for 60% of Tasmania's electricity usage, agreed to a series of voluntary load reductions of up to 100 MW on a sustained basis, contributing to a 1.7% reduction in the output of the manufacturing sector (Eslake, 2016). The total cost of the fires and floods was assessed at

Box 6.1 (continued)

300 million USD. In response funding has been increased to government agencies responsible for managing floods and bushfires, and multiple independent reviews have recommended major policy reforms that are now under consideration (Blake, 2017; Tasmanian Climate Change Office, 2017).

This case illustrates the concepts presented in Figure 6.1. Anthropogenic climate change *likely* contributed to the severity of multiple hazards; including coincident and sequential events (droughts and bushfires, followed by extreme rainfall and floods). Compound risks, including risks for the safety of residents affected by floods and fires, the natural environment affected by MHWs and fires and the economy in the food and energy sectors arose from these climate events with cascading impacts on the industrial sector more broadly as it responded to the shortfall in energy supply.

Extremes experienced in 2015–2016 in Tasmania are projected to become more frequent or more intense due to climate change, including dry springs and summers (Bureau of Meteorology and Australian CSIRO, 2007), intense lows bringing extreme rains and floods in summer (Grose et al. 2012), and MHWs on the east coast associated with convergence of heat linked to the East Australia Current (Oliver et al. 2017) indicating that climate change by increasing the frequency or intensity of multiple climate events will *likely* increase compound risk and cascading impacts (*high confidence*).

Case Study 2: The Coral Triangle

The Coral Triangle is under the combined threats of mean warming, ocean acidification, temperature and sea level variability (often associated with both El Niño and La Niña), coastal development and overfishing, leading to reduced ecosystem services and loss of biodiversity. The Coral Triangle covers 4 million square miles of ocean and coastal waters in Southeast Asia and the Pacific, in the area surrounding Indonesia, Malaysia, Papua New Guinea, the Philippines, Timor Leste and the Solomon Islands. It is the centre of the highest coastal marine biodiversity in the world due to its geological setting, physical environment, and an array of ecological and evolutionary processes which makes it a conservation priority. Together with mangroves and seagrass beds, the 605 species of corals including 15 regional endemics (Veron et al. 2011) provide ecosystem services to over 100 million people from diverse and rich cultures, in particular for food, building materials and coastal protection.

The riches of the ecosystems in the Coral Triangle led to expanding human activities, such as coastal development to accommodate a booming tourism sector and overfishing. There is agreement that these activities, including coastal deforestation, coastal reclamation, destructive fishing methods and over-exploitation of marine life generate important pressures on the ecosystem (Pomeroy et al. 2015; Ferrigno et al. 2016; Huang and Coelho, 2017). As a result, the coastal ecosystems of the Coral Triangle have already lost 40% of their coral reefs and mangroves over the past 40 years (Hoegh-Guldberg et al. 2009).

Risks from compound events include increase in sea surface temperature (SST), SLR and increased human activities. The increasing trend in SSTs was estimated to be 0.1°C per decade between 1960 and 2007 (Kleypas et al. 2015) but increased to 0.2°C per decade from 1985 to 2006 (Penaflor et al. 2009), an estimation comparable with that in the South China Sea (Zuo et al. 2015). However, waters in the northern and eastern parts are warming faster than the rest of the region, and this variability is increased by local parameters linked to the complex bathymetry and oceanography of the region (Kleypas et al. 2015). Areas in the eastern part have experienced more thermal stress events, and these appear to be more likely during La Niña events, which generate heat pulses in the region, leading to bleaching events, some of them already triggered by El Niño Southern Oscillation (ENSO) events. In the Coral Triangle, El Niño events have a relative cooling effect, while La Niña events are accompanied by warming (Penaflor et al. 2009). The 1997–1998 El Niño was followed by a strong La Niña so that degree heating weeks (DHW) values in many parts of the region were greater than four, which caused widespread coral bleaching (DHW values greater than zero indicate there is thermal stress, while DHW values of 4 and greater indicate the existence of sufficient thermal stress to produce significant levels of coral bleaching; Kayanne, 2017). However, in Indonesia, the 2015–2016 El Niño event had impacted shallow water reefs well before high SSTs could trigger any coral bleaching (Ampou et al. 2017). Sea level in Indonesia had been at its lowest in the past 12 years following this El Niño event and this had affected corals living in shallow waters. Substantial mortality was likely caused by higher daily aerial exposure during low tides and warmer SST associated with shallow waters. Another climate change-associated impact in the Coral Triangle is ocean acidification. Although less exposed than other reefs at higher latitudes (van Hooidonk et al. 2013), changes in pH are expected to affect coral calcification (DeCarlo et al. 2017), with an impact on coral reef fisheries (Speers et al. 2016).

At present, different approaches are used to manage the different risks to coral ecosystems in the Coral Triangle such as fisheries management (White et al. 2014) and different conservation initiatives (Beger et al. 2015), including coral larval replenishment (dela Cruz and Harrison, 2017) and the establishment of a region-wide marine protected area system (e.g., Christie et al. 2016).

Box 6.1 (continued)

There is *high confidence* that reefs with high species diversity are more resilient to stress, including bleaching (e.g., Ferrigno et al. 2016; Mellin et al. 2016; Mori, 2016). Sustainable Management of coastal resources, such as marine protected areas is thus a commonly used management approach (White et al. 2014; Christie et al. 2016), supported in some cases by ecosystem modelling projections (Weijerman et al. 2015; Weijerman et al. 2016). Evaluations of these management approaches led to the development of guiding frameworks and supporting tools for coastal area managers (Anthony et al. 2015); however biological and ecological factors are still expected to limit the adaptive capacity of these ecosystems to changes (Mora et al. 2016).

Case Study 3: Severe Atlantic Hurricanes of 2017

The above-average hurricane activity of the 2017 season led to the sequential occurrence of Hurricanes Harvey, Irma and Maria on the Caribbean and southern US coasts (Klotzbach and Bell, 2017), collectively causing 265 billion USD damage and making 2017 the costliest hurricane season on record (Blake et al. 2011; Blake and Zelinsky, 2018).

The role of climate change in contributing to the severity of these recent hurricanes has been much discussed in the public and media. It has not been possible to identify robust long-term trends in either hurricane frequency or strength given the large natural variability, which makes trend detection challenging especially given the opposing influences of greenhouse gases (GHGs) and aerosols on past changes. However, observational data shows a warming of the surface waters of the Gulf of Mexico, and indeed most of the world's oceans, over the past century as human activities have had an increasing impact on our climate (Sobel et al. 2016).

Hurricane Harvey brought unprecedented rainfall to Texas and produced a storm surge that exceeded 2 m in some regions (Shuckburgh et al. 2017). Climate change increased the rainfall intensity associated with Harvey by at least 8% (8–19%; Risser and Wehner, 2017; van Oldenborgh et al. 2017) (*high confidence*). Emanuel (2017) estimated that the annual probability of 500 mm of area-averaged rainfall had increased from 1% in the period 1981–2000 to 6% in 2017. Furthermore, if society were to follow RCP8.5, the probability would increase to 18% over the period 2081–2100.

The event attribution method of Emanuel (2017) indicates that for TC Irma, which impacted the Caribbean islands of Barbuda and Cuba, the annual probability of encountering Irma's peak wind of 160 knots within 300 km of Barbuda increased from 0.13% in the period 1981–2000 to 0.43% by 2017, and will further increase to 1.3% by 2081–2100 assuming RCP8.5. TC Maria followed Irma, and made landfalls on the island of Dominica, Puerto Rico, and Turks and Caicos Islands. The annual probability of encountering Maria's peak wind of 150 knots within 150 km of 17ºN, 64ºW increased from 0.5% during 1981–2000 to 1.7% in 2017, and will increase to 5% by 2081–2100 assuming RCP8.5.

At least 68 people died from the direct effects of Harvey in Houston (Blake and Zelinsky, 2018). The Houston metropolitan area was devastated with the release of about 4.6 million pounds of contaminants from petrochemical plants and refineries. Irma caused 44 direct deaths (Cangialosi et al. 2018) and wiped out housing, schools, fisheries and livestock in Barbuda, Antigua, St. Martin and the British Virgin Islands (ACAPS et al. 2017). Maria caused 31 direct deaths in Dominica, two in Guadeloupe and around 65 in Puerto Rico (Pasch et al. 2018), and completely vacated Barbuda. Maria destroyed almost all power lines, buildings and 80% of crops in Puerto Rico (Rexach et al. 2017; Rosselló, 2017), and damaged pharmaceutical industries that provided 33% of Puerto Rico's gross domestic product (GDP) causing shortages of some medical supplies in the USA (Sacks et al. 2018). The effects of Maria are expected to increase the poverty rate by 14% because of unemployment in tourism and agriculture sectors for more than a year in Dominica (The Government of the Commonwealth of Dominica, 2017), and resulted in outmigration to neighboring countries or the USA (ACAPS et al. 2017; Rosselló, 2017). These economic and social consequences are indicative of the cascading impact of the 2017 hurricanes. The post-disaster reconstruction plan is to renovate telecommunications, develop climate resilient building plans and emergency coordination (Rosselló, 2017; The Government of the Commonwealth of Dominica, 2017).

Collectively, these case studies indicate that climate change has played a role in multiple coincident or sequential extreme events that have led to cascading impacts (*high confidence*). Climate change is projected to increase the frequency or intensity of multiple climate events in the future and this will *likely* increase risks of compound events and cascading impacts (*high confidence*).

6.9 Governance and Policy Options, Risk Management, Including Disaster Risk Reduction and Enhancing Resilience

6.9.1 Decision Making for Abrupt Change and Extreme Events

As outlined earlier in this report, several approaches exist for adaptive responses towards climate change impacts. Other sections that deal with adaptation responses to extremes include Section 1.5.2, Section 4.4 (SLR and coastal flooding), Cross Chapter Box 4 in Chapter 1 and Section 5.5.2.5 in Chapter 5 (adaptation limits for coastal infrastructure and ecosystems). Here, we address adaptation responses especially to abrupt and extreme changes (for responses to special abrupt changes (e.g., AMOC; see also Section 6.7).

Since AR5, growing discussions have advocated for transformative adaptation, implying that they support fundamental societal shift towards sustainability and climate-resilient development pathways (Moloney et al. 2017; IPCC, 2018; Morchain, 2018). Successful adaptation to abrupt change and extreme events incorporates climate change concerns and the impact of climate extremes on vulnerable populations taking into account community participation and local knowledge (Tozier de la Poterie and Baudoin, 2015). These interventions reduce risk and enhance resilience, and contribute to the SDGs and social justice (Mal et al. 2018). Temporal scales denote before and after abrupt changes and extreme events (prevention and post-event response), long- and short-term adaptation measures, and the lag time between forecast, warning and event (Field et al. 2012; IPCC, 2012). Spatial dimensions include local risk management and adaptation as well as regional and international coordination to prepare for unexpected extremes tackling the impacts at multiple geographic scales (Devine-Wright, 2013; Barnett et al. 2014; Lyth et al. 2016; Barange et al. 2018).

Decision making about abrupt change or extreme events is not autonomous; it is constrained by formal and informal institutional processes such as regulatory structures, property rights, as well as culture, traditions and social norms (Field et al. 2012; IPCC, 2012). Efforts in various countries and large cities to improve resilience and adaptation are growing, and these efforts are linked to a global network of research, information and best practices (e.g., Aerts et al. 2014). In both northern and southern high latitudes, extreme climatic conditions and remoteness from densely populated regions constrain human choices. The question is whether responses to extremes and abrupt changes require approaches that are different from the anticipatory management of adaptation to changes in climate and weather extremes. While there are several impact studies on extreme events and abrupt change, very few focus on the necessity of dedicated individual, governmental or business adaptive responses (Tol et al. 2006; Anthoff et al. 2010; Anthoff et al. 2016).

Making appropriate decisions to manage abrupt change and extreme events given deep uncertainty is challenging (Weaver et al. 2013; see Cross-Chapter Boxes 4 and 5 in Chapter 1). This requires the construction of new models integrating different uncertainties under extreme or abrupt scenarios and evaluation of value for money (Weaver et al. 2013). Examples include the inclusion of rapid SLR for assessing coastal impacts and adaptation options (Ranger et al. 2013; Haasnoot et al. 2018; see Sections 6.4 and 6.7). Decision analysis frameworks such as 'Robust Decision Making', 'Decision Scaling', 'Assess Risk of Policy', 'Info-gap', 'Dynamic Adaptation Policy Pathways', 'Dynamic Adaptive Pathways Planning', 'Multi-Criteria Decision Analysis', 'Real Options Analysis' and 'Context-First' accommodate a wide range of uncertainties with subsequent socio-ecological impact (Weaver et al. 2013). The central question remains, however, how one can overcome path dependencies which may cause technical lock-ins in the current system. Monitoring systems of climatic and derived variables, in order to predict necessary shifts in adaptation policies are in development (Haasnoot et al. 2015). However, these frameworks have so far been mostly applied to more gradual shifts of climate change, rather than extreme events and abrupt changes.

Request for the use of 'actionable' information and communication based on climate science and modelling will increase (McNie, 2007; Moser and Boykoff, 2013). Such information can only be effective when it is perceived as 'credible, salient, and legitimate' (Paton, 2007; Paton, 2008; Dilling et al. 2015). Since SREX (IPCC, 2012), there is *medium confidence* that trust in the information and the institution (Hardin, 2002; Townley and Garfield, 2013) that governs extreme events and abrupt change (Malka et al. 2009; Birkmann et al. 2011; Schoenefeld and McCauley, 2016) is important. Trust in expert and scientific knowledge helps people make sense of climate change impact and engage with adaptation measures (Moser and Boykoff, 2013; Yeh, 2016). Without such knowledge, people have little recourse to believe and evaluate relevant information (Bråten et al. 2011). Individuals who trust their government can be complacent and do not prepare for the consequences of extremes (Simpson, 2012; Edmondson and Levy, 2019), and shift the responsibility to the government (Edmondson and Levy, 2019). Familiarity with and information about hazards, community characteristics, as well as the relationship between people and government agencies influence the level of trust (Paton, 2007).

Recent literature shows that there are crucial differences between the ethical challenges of mitigation and those of adaptation (Wallimann-Helmer, 2015; Wallimann-Helmer, 2016) in their dealings with Loss and Damage (L&D); and the ongoing analysis disputes how to distribute responsibilities between mitigation and adaptation based on climate justice criteria (Wallimann-Helmer et al. 2019). The Warsaw International Mechanism on L&D under the United Nations Framework Convention on Climate Change (UNFCCC) addresses irreversible changes and limits to adaptation at the global scale (see also Cross-Chapter Box 1 in Chapter 1). This is in contrast to national and local policies, addressing impacts and adaptation. Within the SROCC report, several of the documented and projected irreversible or unavoidable and thus residual impacts beyond adaptation would potentially fall under this category (e.g., Warner and van der Geest, 2013; Huggel et al. 2019; Mechler et al. 2019), including impacts from SLR, land erosion and reduced freshwater resources on small islands, changes in high mountains and cryosphere changes, as well as changes in ocean species and resources. Apart from climate hazards, risks for L&D are also determined by increasing exposure and vulnerability

(Birkmann and Welle, 2015). Such impacts can be assessed using conventional frameworks, but the debate on the precise scope of such impacts remains, including those from anthropogenic climate change impacts as well as natural climate variability and extremes (e.g., James et al. 2014). More work is required to explore the range of activities available for responding to L&D resulting from slow onset processes in the scope of the SROCC report such as ocean acidification (Harrould-Kolieb and Hoegh-Guldberg, 2019) and mountain cryosphere changes (Huggel et al. 2019).

Under the same L&D mechanism, risk transfer mechanisms and insurance have been suggested as a specific adaptation policy option. Several forms of 'climate change' insurance have been proposed recently, but their potential for adaptation has met with criticism, importantly because of the costs of formal insurance and other risk transfer options, as well as issues with sustainability given the lack of loss prevention and adaptation (Surminski et al. 2016; Linnerooth-Bayer et al. 2019). A compensation mechanism for low-lying small islands inclusive of L&D proposal is in progress (Adelman, 2016). Insurance (see also Section 4.4.4) can help absorb extreme shocks for both individuals, using traditional insurance and parametric insurance. Sovereign insurance mechanisms can help governments absorb large losses (Linnerooth-Bayer et al. 2019), but eventually they need to be coupled with other incentives for adaptation and risk reduction measures to be cost-effective (Botzen, 2013) (*medium confidence*).

There is a consensus that investing in disaster risk reduction has economic benefits, although there is *medium evidence* about the range of the estimated benefits which varies from a global estimate of two to four dollars saved for each dollar invested (Kull et al. 2013; Mechler, 2016) to about 400 EUR per invested 1 EUR in the case of flood early warning systems in Europe (Pappenberger et al. 2015). The US Federal Emergency Management Agency indicated that a 1% increase in annual investment in flood management decreases flood damage by 2.1% (Davlasheridze et al. 2017). Conserving ecosystems that provide services for risk reduction also has monetary benefits. Wetlands have been observed to reduce damages during storms. Wetlands and floodplains in Otter Creek (Vermont, USA) reduced damages caused by storms by 54–78% and 84–95%, respectively, for Tropical Storm Irene (Watson et al. 2016). For the whole of the USA, wetlands provide 23.2 billion USD yr^{-1} in storm protection services and the loss of 1 hectare of wetland is estimated to correspond to an average 33,000 USD increase in storm damage from specific storms (Costanza et al. 2008). Engineered structures are also expected to reduce risks. In Europe, to maintain the coastal flood loss constant relative to the size of the economy, flood defence structures need to be able to protect coastal areas for a projected increase of sea level between 0.5–2.5 m. Without these risk reduction actions, the expected damages from coastal floods could increase by two or three degrees of magnitude compared to the present (Vousdoukas et al. 2018). Although risk reduction actions are generally considered an effective way to reduce the damages by shifting the loss-exceedance curve, cost-benefit analysis of disaster risk reduction actions faces several challenges, including its limited role in informing decisions, spatial and temporal uncertainty scales, and discounting and choice of discount rate that affect cost-benefit analysis results heavily (Mechler, 2016).

6.9.2 Transformative Governance and Integrating Disaster Risk Reduction and Climate Change Adaptation

Governance for effective adaptation defined as changes in practice, process and structure (Smit et al. 2001) considers equity, legitimacy and co-benefits (Patterson et al. 2018) appropriate to the issue (Young, 2002). Countries, sectors and localities place different values and perspectives on these categories, and they can change over time (Plummer et al. 2017; see Cross-Chapter Boxes 1 and 2 in Chapter 1). Transformative governance embraces a wider application of climate change-induced mitigation and adaptation strategies to generate fundamental change. It is society-wide and goes beyond the goals of climate change policies and measures (IPCC, 2013; Patterson et al. 2018). It is distinguished from conventional strategies and solutions, as it includes both natural and human systems and intertwines with the SDGs (Fleurbaey et al. 2014; Tàbara et al. 2019). Transformational adaptation is also needed when incremental adaptation to extreme events and abrupt changes is insufficient (Kates et al. 2012). Planned retreat from SLR and climate refugees illustrate the need for transformative governance as the current coastal and risk management regimes do not have the capacity to handle these issues adequately. Inclusion of bilateral and regional agreements related to climate-induced migration (McAdam, 2011), land use planning frameworks to respond to policy, institutional and cultural implications of migration (Matthews and Potts, 2018), and identification of beneficiaries of managed retreat (Hino et al. 2017) along with positive opportunities for migrants to diversify income and avoid being in harm's way (Gemenne, 2015) are steps towards transformative governance. Retreat and migration entail local responses that include indigenous and local knowledges and perspectives that can be applied to solve these issues (Farbotko and Lazrus, 2012; Hilhorst et al. 2015; Tharakan, 2015; Iloka, 2016; Nunn et al. 2016; see also Cross-Chapter Boxes 2 and 5 in Chapter 1). Another example is the Polar region which has started to pursue transformative governance given the potential for increased tourism and cooperation that require changed governance structure (see Sections 3.5.2; 3.5.5 and Table 3.7 in Chapter 3). Accountability for transformations and transitions has been identified as a crucial factor to support responsible action and strengthen climate governance (Edmondson and Levy, 2019).

Though discourse abounds related to transformative governance, it falls short of its ideal in climate change action plans as it is unclear whether communities have the capacity to engage in substantive change to build low-carbon and resilient communities (Burch et al. 2014). The results of a study on the USA by Tang and Dessai (2012) indicate that climate adaptation and mitigation plans' treatment of extreme climate conditions and disaster preparedness is limited. Moreover, risk communication with the public is part of an integrated disaster warning system, but behavioural response to disaster warnings are often governed by personal beliefs about the nature of the hazard; and ultimately swaying individual decisions to comply with or ignore the warning message (Mayhorn and McLaughlin, 2014). New approaches such as the 'first mile' of early warning systems, built on the specific needs from beneficiary communities instead of on technological progress, are being implemented (Zommers et al. 2017); but they have not yet been assessed.

Coupling disaster risk reduction and management with climate change adaptation effort – following the set targets of UNFCCC and the Sendai Framework – has shown progress since SREX and AR5 (e.g., Lawrence and Saunders, 2017). Substantial literature exists on the topic, but there is little assessment of practices on the ground in the implementation of integrated disaster management and climate change adaptation (Nalau et al. 2016) including health (Banwell et al. 2018). Mainstreaming disaster risk reduction and climate change adaptation within and across sectors is considered essential to ensure administrative coordination and coherence across sectoral plans and policies (Shimizu and Clark, 2015) (*medium confidence*). Financial and technological support and capacity building especially related to public works, savings or loans enable households to build assets and improve livelihoods (Ulrichs et al. 2019). No assessment is available so far of the efficiency and effectiveness of mainstreaming especially related to the integration of climate change adaptation and disaster risk reduction, let alone for abrupt and extreme impacts.

Case studies of integration note major problems, for example, weak coordination among government agencies (Seidler et al. 2018); lack of data and user-friendly information to guide decision making at the local level (Jones et al. 2017) and the need for the central governmental support for data availability (Putra et al. 2018); fragmentation due to competing local objectives (Forino et al. 2017); dependence on regional and international frameworks in the absence of a national framework (Rivera and Wamsler, 2014); limited availability of formal training in integration (Hemstock et al. 2017); and turf wars between responsible government agencies (Nemakonde and Van Niekerk, 2017). The case of Pacific islands such as Vanuatu is indicative of these problems. Though they have coupled disaster risk reduction with climate change adaptation, problems manifest in relationships, responsibilities, capacity and expectations between government agencies and other actors (e.g., international donors and non-governmental organisations), as analysed by Vanuatu's response to the Category 5 TC Pam (Nalau et al. 2017). Some solutions are proposed such as getting all the actors on the same page and focusing on reducing vulnerability to longer-term environmental hazards (Schipper et al. 2016); focussing on specific goals, objectives and strategies (Organization of American States, 2014); assigning a single department to handle integration (APEC, 2016); and citing real-life decision examples in national guidelines (Bell et al. 2017). Place-based responses also entail the inclusion and the acknowledgement of indigenous and local knowledge for an enhanced resilience pathway (Hilhorst et al. 2015; Tharakan, 2015; Iloka, 2016; Nunn et al. 2016).

Given the significance of disaster risk reduction to enhance climate change adaptation regardless of the integration of the two, the Sendai Framework for Disaster Risk Reduction 2015–2030 focuses on seven targets and four priorities that foster participation beyond information sharing and include partnerships and collaborations within society (UNISDR, 2015). Inclusion of, and coordination between, different stakeholders is a key component for managing risks of extreme events, including in a changing climate (*medium confidence*). In the Wadden Sea coastal area, for example, crucial parts of coordinating disaster risk reduction, include (i) responsibility-sharing among authorities, sectors and stakeholders, (ii) all-of-society engagement and partnership with empowerment and inclusive participation, and (iii) development of international, regional, subregional and transboundary cooperation schemes (González-Riancho et al. 2017). In India, a change in the coordination structure was pivotal in reducing fatalities from over 10,000 to 45 between cyclones Orissa (Odisha) in 1999 and Phailin in 2013. In this case, the Disaster Management Act of 2005 established a comprehensive policy and command and control system during disaster response that empowered the most qualified government officials regardless of their rank. This system provides authority to and holds accountability for those in charge of ground operations. Though this rigid system may sometimes be questioned, a unified and top-down command structure works better when there is a lack of mature disaster management system (Pal et al. 2017).

In sum, limiting the risk from the impact of extreme events and abrupt changes leads to successful adaptation to climate change if climate-affected sectors and disaster management relevant agencies coordinate well (*high confidence*). Transformative governance, including successful integration of disaster risk management and climate change adaptation, empowerment of vulnerable groups, accountability of governmental decisions, and longer-term planning promotes climate-resilient development pathways (*high confidence*). An enhanced understanding of the institutional capacity as well as the legal framework addressing abrupt changes and extreme events is especially important (*medium confidence*).

Knowledge gaps limit the identification of the most relevant actions to achieve and pursue climate-resilient development pathways. Since SREX and AR5, there is little research on indirect impacts of climatic extremes on ecosystems and consequences on poverty and livelihoods critical to the SDGs. For example, adaptation solutions and limitations, including governance challenges, for the ocean do not include extreme events (Sections 5.5.2 and 5.5.3 in Chapter 5). Further, there is only scant literature on L&D, including non-economic impacts, resulting from well-documented processes such as MHWs (Section 6.4), SLR impacts on low-lying coasts (Section 4.3), and cryosphere changes (Section 2.3; Chapter 3) (*high confidence*). Limited information is available concerning the cost-benefit and effectiveness of risk-reduction measures. Coupling risk transfer and insurance mechanisms with risk reduction measures, for example, can enhance the cost-effectiveness of adapting to climate change (*medium confidence*).

Frequently Asked Questions

FAQ 6.1 | How can risks of abrupt changes in the ocean and cryosphere related to climate change be addressed?

Reducing greenhouse gas (GHG) emissions will reduce the occurrence of extreme events and the likelihood of abrupt changes. Abrupt changes can be irreversible on human time scales and, as tipping points, bring natural systems to novel conditions. To reduce risks that emerge from these impacts of climate change, communities can protect themselves or accommodate to the new environment. In the last resort, they may retreat from exposed areas. Governance that builds on diverse expertise and considers a variety of actions is best equipped to manage remaining risks.

Climate change is projected to influence extreme events and to potentially cause abrupt changes in the ocean and the cryosphere. Both these phenomena can add to the other, slow-onset impacts of climate change, such as a global warming or sea level rise (SLR). In addition, abrupt changes can be tipping points, bringing the ocean, cryosphere, as well as their ecosystems, or the whole climate system, to new conditions instead or going back to the ones prevailing before the abrupt change.

In the ocean, a possible abrupt change is associated with an interruption of the Atlantic Meridional Overturning Circulation (AMOC), an important component of global ocean circulation. A slowdown of the AMOC could have consequences around the world: rainfall in the Sahel region could reduce, hampering crop production; the summer monsoon in Asia could weaken; regional SLR could increase around the Atlantic, and there might be more winter storms in Europe. The collapse of the West Antarctic Ice Sheet (WAIS) is considered to be one of the tipping points for the global climate. Such an event can be triggered when ice shelves break and ice flows towards the ocean. While, in general, it is difficult to assess the probability of occurrence of abrupt climate events they are physically plausible events that could cause large impacts on ecosystems and societies and may be irreversible.

Reducing GHG emissions is the main action to limit global warming to acceptable levels and reduce the occurrence of extreme events and abrupt changes. However, in addition to mitigation, a variety of measures and risk management strategies supports adaptation to future risks. Future risks linked to abrupt changes are strongly influenced by local conditions and different characteristics of the events themselves and evolve differently depending on the circumstances. One major factor for adaptation is whether the extreme events will simply amplify the known impacts or whether they will cause completely new conditions, which may be related to a tipping point. Another essential factor is whether an extreme event or abrupt change will happen in isolation or in conjunction with other events, in a chain of cascading impacts or as part of a compound risk where several events happen at the same time so that impacts can multiply each other. Also, impacts are heavily aggravated by increasing exposure and changes in vulnerability, for example reducing the availability of food, water and energy supply, and not just the occurrence of extremes themselves.

Successful management of extreme events and abrupt changes in the ocean and cryosphere involves all available resources and governance approaches, including among others land-use and spatial planning, indigenous knowledge and local knowledge. The management of the risks to ecosystems include their preservation, the sustainable use of resources and the recognition of the value of ecosystem services. There are three general approaches that, alone or in combination, can enable communities to adapt to these events: retreat from the area, accommodation to new conditions and protection. All have advantages and limitations and their success will depend on the specific circumstances and the community's level of adaptability. But only transformative governance that integrates a variety of strategies and benefits from institutional change helps to address larger risks posed by compound events. Integrating risk-reduction approaches into institutional practices and inclusive decision making that builds on the respective competences of different government agencies and other stakeholders can support management of these extremes. A change of lifestyles and livelihoods might further support the adaptation to new conditions.

6.10 Knowledge Gaps

A comprehensive, detailed list of all the knowledge gaps that have been identified during the assessment performed in this chapter is not possible, hence we focus here on gaps that are relevant for multiple phenomena.

Detection, attribution and projection of physical aspects of climate change at regional and local scales are generally limited by uncertainties in the response of climate models to changes in GHGs and other forcing agents. Additionally, regionally-based attribution studies for extreme events may be lacking in some areas, possibly reflecting the lack of capacity or imperative by regional and national technical institutions to undertake such studies. Thermodynamic aspects of change may be more robust than those involving changes in dynamics e.g., the tracks of TCs or ocean dynamical components of MHW formation. Increasing resolution and improvements in climate models may help to reduce uncertainty. However, because extreme events and highly nonlinear changes (e.g., AMOC collapse) are, by definition, found in the 'tails' of distributions, ensembles or long climate model runs may be required.

While it may not be possible to quantify the likelihood of very rare events or irreversible phenomena, it may be possible to quantify their impacts on natural and human systems. Such information may be more useful to policy makers (Sutton, 2018). Impacts on natural systems (e.g., marine ecosystems) are in general better quantified than impacts on human systems, but there are still many gaps in the literature for the phenomena assessed here (e.g., future impacts of extreme El Niño and La Niña events). The body of literature on compound risks and cascading impacts is growing but is still rather small. One area where there seems to be a serious lack of literature is in the assessment of the economic impacts of extreme and abrupt/irreversible events.

Literature on managing risks and adaptation strategies for abrupt and irreversible events is sparse, as is the literature on the combined impacts of climate-driven events and societal development or maladaptation. The same is true for compound risks and cascading impacts. Theory on transformative governance is emerging but practical demonstrations are few.

Finally, research is still often 'siloed' in physical modelling, ecosystem modelling, social sciences etc. Researchers who can cross boundaries between these disciplines will help accelerate research in the areas covered by this chapter.

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