Chapter 3: Polar Regions

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Date of Draft: 16 November 2018

Notes: TSU Compiled Version

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Executive Summary

Despite their locations at the opposing ends of the planet, the polar regions are interconnected parts of the Earth System that exert significant influence over the lives and livelihoods of humanity via shared ocean, atmosphere, ecological and social systems. This chapter assesses the state of interdisciplinary knowledge concerning the different key elements of the Arctic and Antarctic systems, how they are affected by climate change and how they are likely to develop in the future. Concurrently, it assesses the local, regional and global consequences and impacts of such changes, and the opportunities and challenges of different response options. Key findings from this chapter are as follows.

Why do the polar regions matter, regionally and globally, and how are they changing?

Climate-induced changes to the cryosphere and ocean in the polar regions have global consequences and impacts that are evident now (*very high confidence*[1]). There is strong evidence that the pervasive regional changes in sea ice, seasonal snow, ice sheets, permafrost and ocean that have been observed in the polar regions have consequences and impacts across the globe via atmospheric, marine and economic linkages. {Box 3.1, 3.2.1, 3.2.3, 3.2.4, 3.3.1, 3.3.3, 3.4.1, 3.4.3}

The coupled Arctic ocean/cryosphere system is now in a markedly different state than at the end of the 20th century (*very high confidence*). Evidence for this state change, with increases in surface temperature at approximately twice the rate of the global average (*very high confidence*), is derived from multiple individual and linked Arctic regional changes and their consequences and impacts. {Box 3.1, 3.2.1, 3.3.1, 3.3.2, 3.4.1}

The polar oceans are changing more rapidly than the global ocean as a whole, with consequences for climate and ecosystem services (*high confidence*). The amounts of heat and carbon stored in the polar oceans have increased in recent decades, with marked ocean warming in both polar regions and reinvigoration of the Southern Ocean carbon sink since the early 2000s. These processes modify the rates of global climatic change, and have impacts on regional marine ecosystems via ocean temperature change and acidification. Increased heat uptake by the Southern Ocean has been attributed to anthropogenic processes, most notably the influence of greenhouse gases. {3.2.1}

Climate-induced changes in both polar oceans, sea ice, and the terrestrial cryosphere drive shifts in habitats that affect the ranges and abundance of ecologically important species that are of global commercial and conservation value (*high confidence*). This includes habitat expansion of several boreal fish and crab stocks in the Barents Sea in the European Arctic that are commercially exploited (*high confidence*). Climate projections indicate further shifts in the future, including habitat contraction for Antarctic krill, a keystone species in Southern Ocean foodwebs that is the focus of an international fishery, and loss of Antarctic seafloor biodiversity (*medium confidence*). On Arctic land, projections indicate a loss of globally unique biodiversity as some high-Arctic species will be outcompeted by more southerly species and very limited refugia exist (*medium confidence)*. Projected range expansion of subarctic marine species will increase competition pressure for high-Arctic species (*medium confidence*), with regionally-variable impacts dependent on physical and ecological conditions. {3.2.3; Box 3.3}

Substantial declines in Arctic summer sea ice and spring snow cover extent observed over the period of satellite measurements have continued over the past decade (*high confidence*), with consequences for the global climate system. Observed and projected reductions in snow extent and sea ice extent and thickness affect the global climate on annual to decadal time scales via sustained albedo decreases affecting surface energy budget and energy absorption (*very high confidence*). Emerging evidence indicates that changes in Arctic sea ice can influence weather outside the Arctic on timescales of weeks to month (*low confidence*). {3.3.1.1; 3.4.1.1; Box 3.1}

[1] FOOTNOTE: In this Report, the following summary terms are used to describe the available evidence: limited, medium, or robust; and for the degree of agreement: low, medium, or high. A level of confidence is expressed using five qualifiers: very low, low, medium, high, and very high, and typeset in italics, e.g., *medium confidence*. For a given evidence and agreement statement, different confidence levels can be assigned, but increasing levels of evidence and degrees of agreement are correlated with increasing confidence (see Section 1.9.3 and Figure 1.4 for more details).

Permafrost temperatures have continued to increase to record high levels (*high confidence*); that trend is projected to continue, with consequences for the global climate system. The organic carbon pool stored in Arctic and boreal permafrost zone soils contains almost twice the carbon in the atmosphere (*high confidence*). Changes in permafrost influence global climate through emissions of the greenhouse gases carbon dioxide and methane released from the microbial breakdown of organic carbon. Evidence suggests substantial loss of permafrost carbon to the atmosphere by 2100 and beyond under weak mitigation, while scenarios limiting anthropogenic carbon emissions will result in lower losses (*high confidence*). There is *low confidence* concerning the level to which increased plant growth will compensate these losses. {3.4.1; 3.4.2; 3.4.3}

Since around 2000, it is *virtually certain* that the Greenland ice sheet has lost mass, and *very likely*[2] that the Antarctic ice sheet has lost mass. The rate of mass loss of the Greenland Ice Sheet and polar glaciers has increased since around the year ~2000 (*high confidence*). The rate of Antarctic Ice Sheet overall mass loss has increased since the mid-2000s (*medium confidence*), dominated by regions of West Antarctica (*high confidence*). Because of a lack of long-term mass change observations in both polar regions and incomplete representation of the full range of relevant processes in ice sheet models, unambiguous attribution of mass loss from ice sheets to anthropogenic influence is currently not available. {3.3.1, 3.3.2}

Rapid thinning, retreat and acceleration of outlet glaciers is occurring in regions of West Antarctica under the influence of warm ocean waters (*high confidence*); this demonstrates the potential for accelerated rates of future ice sheet loss and sea level rise.New evidence indicates that Antarctic Ice Sheet mass loss is driven predominantly by ocean-induced under-ice shelf melting, enhanced glacier flow and grounding line retreat (*high confidence*). It is not currently clear whether unstable retreat of the West Antarctic ice sheet is underway, and there is potential for accelerated rates and a high magnitude of future sea level rise (*medium confidence*). Greenland Ice Sheet and polar glacier ice losses are dominated by atmosphere-induced surface melt (*high confidence*), limiting their potential to cause large increases in the projected rate of future sea level rise (*medium confidence*). {3.3.1, 3.3.2, Cross-Chapter Box 6, Chapter 4}

What are the impacts and risks of the observed and projected changes and who will be affected?

Projected warming will result in continued loss of Arctic sea ice and terrestrial snow, changes to permafrost, and reductions in the mass of glaciers, but important differences in the trajectories of change emerge from mid-century through end of century depending on mitigation measures that are taken (*high confidence*). Declines in Arctic sea ice extent are projected for all seasons to end of century (*high confidence*). For stabilized global warming of 1.5°C, sea ice free summers in the Arctic are projected to be infrequent; under weak mitigation, the consistent occurrence of ice free summers is expected (*high confidence*). The potential for stabilization of Arctic autumn and spring snow cover losses by mid-century under strong and medium climate change mitigation scenarios contrasts with continued loss to end of century under weak mitigation scenarios (*high confidence*). The magnitude of the projected loss of near-surface permafrost is sensitive to the mitigation scenario (*medium confidence*). Mass loss of Arctic glaciers will be greater under RCP8.5 than RCP2.6 (*medium confidence*). {3.2.2; 3.3.2; 3.4.2}

Climate-induced changes in the polar oceans and cryosphere are altering marine primary production, with impacts on marine foodwebs and ecosystems (*high confidence*). In the Arctic, changes in the timing, duration and intensity of primary production are affecting secondary production, with consequences for species composition, spatial distribution, abundance of higher trophic levels (zooplankton, fish, crustaceans and top predators), and impacts on ecosystem structure and biodiversity {3.2.1; 3.2.3, 3.2.4}. In the Antarctic, primary production is projected to increase in regions near to the Antarctic continent, but the implications for higher trophic levels and for carbon export are not yet determined. {3.2.1; 3.2.3; 3.3.3}

The cascading effects of climate-induced stressors on polar marine ecosystems will have impacts on fisheries, but future risks for linked human systems depend on the level of mitigation and especially

[2] FOOTNOTE: In this Report, the following terms have been used to indicate the assessed likelihood of an outcome or a result: Virtually certain 99–100% probability, Very likely 90–100%, Likely 66–100%, About as likely as not 33–66%, Unlikely 0–33%, Very unlikely 0–10%, Exceptionally unlikely 0–1%. Additional terms (Extremely likely: 95–100%, More likely than not >50–100%, and Extremely unlikely 0–5%) may also be used when appropriate. Assessed likelihood is typeset in italics, e.g., *very likely* (see Section 1.9.3 and Figure 1.4 for more details).

the responsiveness of precautionary management approaches (*medium confidence*). Impacts on fisheries in polar regions will have measurable economic and social implications for regional renewable resource economies, cultures and the global supply of fish and shellfish, including krill. Specific impacts will depend on the degree of climate change and on the strategies employed to manage the effects on stocks and ecosystems. Some current management strategies may not sustain viable commercial fisheries under higher emission scenarios. This exemplifies the limits to the ability of existing natural resource management frameworks to address ecosystem change. {3.2.4; 3.5.3; 3.5.4}

Climate-related reductions in snow and freshwater ice and changes to permafrost on Arctic land are affecting hydrology, disturbance regimes and vegetation, and thereby decreasing water and food security for people (*high confidence)*. These changes influence access to, and food availability within, hunting, fishing, forage and gathering areas, and affect the abundance and distribution of culturally and economically important species such as reindeer (*high confidence*); there are impacts on health and cultural identity of Arctic peoples. Freshwater ecosystems, including fish for harvest, are impacted by changes in surface water conditions and lake ice regimes. There are limits to the success of adaptation measures, which can constrain benefits from new opportunities for subsistence activities arising from ecosystem change. {3.4.1; 3.4.2; 3.4.2; 3.4.3; 3.5.3}

Permafrost change will continue to impact infrastructure in urban and rural areas as well as distributed infrastructure for resource extraction and transportation (*high confidence*). Under weak climate mitigation scenarios, 70% of Arctic circumpolar infrastructure is located in areas where permafrost is projected to thaw by 2050 (*high confidence*). Basing infrastructure design requirements and codes on past environmental records is not sufficient in a changing climate. {3.4.3.3}

Reduction in Arctic sea ice extent and the shift to predominantly seasonal ice cover is leading to new opportunities for marine transport and tourism (*very high confidence*). Arctic ship traffic has increased over the past decade due to enhanced access from changes to sea ice cover. More intense Arctic marine transportation and tourism have significant socio-economic and political implications for global trade, northern nations, and economies strongly linked to traditional shipping corridors; they also create risks for the polar environment and coastal communities because implementation of region-specific regulatory systems is not keeping pace. {3.2.1.1; 3.2.4.2; 3.2.4.3}

What are the options for responding to polar change that reduce risk and support resilience?

Limited knowledge, financial resources, human capital and organisational capacity are constraining adaptation in many human sectors of polar regions (*high confidence*). Harvesters of renewable natural resources are adjusting timing of activities to changes in seasonality and less safe ice travel conditions, municipalities and industry are addressing infrastructure failures associated with flooding and thawing permafrost, and coastal communities and cooperating agencies are now planning for relocation. In spite of these adaptations, many groups are making decisions without adequate knowledge to forecast near- and long-term conditions, and the funding, skills and organizational support to engage fully in planning processes (*high confidence*). {3.5.3, 3.5.5, Cross-Chapter Box 7}

Human responses to climate change in polar regions are in many cases reducing immediate risks with short-term adaptation focused on specific problems, but not building resilience to known future impacts and surprises. The current emphasis on short-term adaptation to specific problems is not sufficient to plan for long-term resilience to the scale, complexity and uncertainty of climate change, and will ultimately not succeed in reducing the risks and vulnerabilities to society. Moving toward a dual focus will require transformation of many institutions, economies and values (*high confidence*).

Innovative approaches to problem solving are better suited to meet the novel challenges of climate change in polar regions by building resilience and supporting sustainable development (*medium confidence*). Assessing, implementing, and continually refining systems of governance that ready society for the projected trajectories, impacts and inevitable surprises of climate change require the attention of decision-makers. Observation systems that draw on a diversity of knowledge from multiple scales, participatory processes for analysis and decision making, and adaptive governance are emerging innovations to meet these challenges. {3.5.2, 3.5.5}

The capacity of polar governance systems to respond to climate change has strengthened recently, but their development is not sufficiently rapid to address the challenges of ongoing and projected changes that pose large risks for societies (*high confidence*). Human responses to climate change in the polar regions occur in a fragmented, multilevel governance landscape that is challenged to address cascading risks and uncertainty in an integrated and precautionary way (*high confidence)*. Simultaneously, climate change and new polar interests from outside the regions are driving stronger coordination and integration between different levels and sectors of governance. These responses modify the cooperation and balance of interests between states and international groups, with informal organisations playing an increasingly active role in shaping climate-change relevant regulations. {3.5.4}

Synthesis

This chapter provides strong evidence of many substantial changes in the Arctic and Antarctic since AR5, with several important new changes detected. Many of these have consequences for human populations across the globe, including via sea level rise, climate feedbacks, and impacts on commercial and industrial operations. Knowledge and observations of the polar regions are sparse compared with many other regions, due to their remoteness from major population centres and the challenges operating within them; Indigenous Knowledge and local knowledge in the Arctic is thus disproportionately valuable when considered in addition to scientific data. Projections of polar systems indicate potential future changes that will require management at the regional level and mitigation at the global level to constrain their consequences and impacts. Strengthened cooperation in observing, understanding and responding to polar changes and their impacts can serve as an exemplar for developing climate resilient pathways globally.

3.1 Introduction: Polar Regions, People and the Planet

Our understanding of the consequences of global climate change for the polar regions continues to broaden and deepen, motivated not least by a growing appreciation of the importance of these regions to planetary systems and to the lives and livelihoods of people across the globe.

Since the IPCC's Fifth Assessment Report (AR5), there has been a growing body of scientific literature, assessments and overviews pertaining to the polar regions. These have afforded improved understanding of the dynamics and functioning of the polar regions in the context of climate change, and offer new knowledge that has the potential to help societies identify responses to ongoing and future changes in the ocean and cryosphere.

The goal of this chapter is to assess the scientific information published since AR5, with a focus on determining the extent to which this new knowledge has changed our understanding of the causes and consequences of polar change, and of how people in polar regions and beyond can respond. To achieve this goal, the chapter provides an integrated assessment across the physical, biological and human dimensions of the polar regions, and considers together the relevant material that in previous IPCC reports would have been assessed separately. This offers the opportunity, for the first time in a global report, to trace cause and consequence through the different polar components of the ocean and cryosphere systems to the point at which biological and social impacts and risks can be determined and related to adaptation options and limits, and responses to enhance resilience.

The polar regions are two integrated parts of the Earth System and interact with the rest of the world through shared ocean, atmosphere, ecological and social systems; notably, they play key roles as important components of the global climate system. Important differences in the physical setting of the two polar regions—the Arctic an ocean surrounded by land, the Antarctic a continent surrounded by an ocean—structure the nature and magnitude of these interactions. This chapter therefore takes a systems approach to assessing cryosphere and ocean changes in the polar regions and emphasises their linkages to the rest of the world in order to better address the key issues of climatic change for the polar regions, the planet and its people (Figure 3.1).

Of equal significance for this chapter is acknowledging the existence of multiple and diverse perspectives of the polar regions, many of them overlapping. For the northern polar region, these multiple perspectives encompass the Arctic as a homeland, a source of resources, a key part of the global climate system, a place for preserving intact ecosystems, and a place for international cooperation. Many of these perspectives are equally relevant for the Antarctic, though with some notable differences, the most significant of which is that the Arctic has a population for whom the region is home. When assessing knowledge relating to climate change in the context of adaptation options and enhancing resilience (see Cross-Chapter Box 1 in Chapter 1), such differences are important as they are linked to diverse human values, social processes, and use of resources.

Consideration of all peer-reviewed scientific knowledge is a hallmark of the IPCC assessment process. Lately, there is increasing awareness of the value of considering in parallel Indigenous Knowledge and local knowledge in integrated assessments of climate change, specifically because the 'multiple ways of knowing' not only broaden and strengthen the knowledge base but also facilitate better understanding of the challenges facing Indigenous Peoples, and identification and acceptance of adaptation strategies in communities across the region (see Cross-Chapter Box 3 in Chapter 1). By incorporating published Indigenous Knowledge and local knowledge in parallel with scientific knowledge, this chapter seeks to demonstrate the benefits of incorporating the multiple ways of knowing for assessing climate change impacts and responses.

There is great complexity within the interdisciplinary understanding of the polar regions, with multiple and often interacting drivers and feedbacks causing diverse, multi-faceted responses that influence physical, biological and human systems. These are outlined in detail throughout the course of this chapter; to help navigation, Figure 3.1 includes pointers to the relevant chapter sections.

Reflecting the global connectivity of the polar regions, we purposefully adopt a flexible approach when describing their spatial footprint in relation to particular subjects or scientific disciplines. Our broad conception is that the southern polar region encompasses the flow of the Antarctic Circumpolar Current at least as far north as the Subantarctic Front and fully enclosing the Convention for the Conservation of Antarctic Marine Living Resources Statistical Areas (CCAMLR, 2017c), whilst the marine Arctic comprises the areas of the Arctic Large Marine Ecosystems (PAME, 2013). The terrestrial Arctic includes the areas of the northern continuous and discontinuous permafrost zone, the Arctic biome, and the parts of the boreal biome that are characterised by cryosphere elements, such as permafrost and persistent winter season snow cover[3]. The spatial footprints of these polar regions (see inset maps in Figure 3.1) include a vast share of the world's ocean and cryosphere: they encompass surface areas equalling 20% of the global ocean and more than 90% of the world's continuous and discontinuous permafrost area, both of the world's ice sheets, 69% of the world's glacier area, almost all of the world's sea ice, and land areas that are entirely snow-covered during winter.

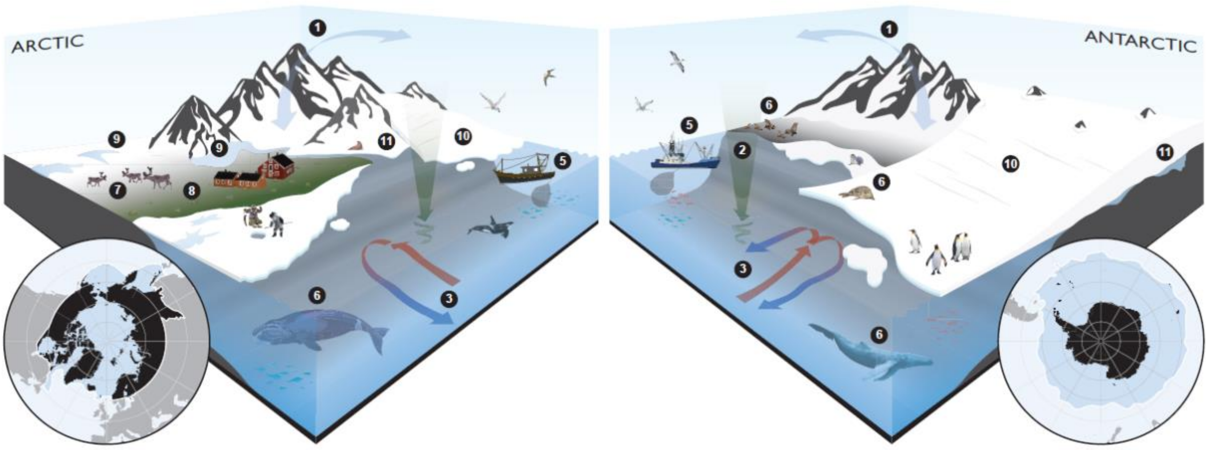


Figure 3.1: Schematic of some of the key features and mechanisms assessed in this Chapter, and by which the cryosphere and ocean in the polar regions influence climate, ecological and social systems in the regions and across the globe. The relevant sections wherein information can be found in this chapter are numbered.
(1) The changing cryosphere influences albedo and atmospheric feedbacks, with global-scale consequences for climate (Box 3.1; Section 3.4.3.1; Appendix 3.A.1);
(2) The polar oceans are key regions for the drawdown and storage of heat and carbon (including anthropogenic) from the atmosphere (Section 3.2);
(3) Processes in the polar oceans exert strong influences on water mass formation, and driving/closure of the global ocean circulation (Section 3.2);
(4) The Arctic is home to local and Indigenous populations, whose daily life and rich and diverse cultural heritage is closely intertwined with the cryosphere (Sections 3.2.4; 3.4.3.3, 3.5);
(5) The polar regions are of increasing economic significance, bringing risks and opportunities and new challenges to cooperation, governance, and development (Sections 3.2.4, 3.4.3.3, 3.5);
(6) The polar oceans are key sites for marine biodiversity and ecosystems, with some species subject to globally-relevant commercial exploitation (Sections 3.2.3; 3.5)
(7) The polar terrestrial regions feature unique biodiversity that is affected by changes in climate and the cryosphere, with impacts on people (Section 3.4.3);
(8) Changing snow and frozen ground affects Arctic landscapes, with consequences for plants, wildlife, ecosystems, people, and global climate (Section 3.4);
(9) Terrestrial freshwater systems influence hydrological and ecological processes on land and off shore, with impacts on northern populations (Section 3.4);
(10) Meltwater discharged from the Greenland and Antarctic Ice Sheets exerts influences on global sea level, ocean stratification and circulation (Section 3.3)
(11) Subglacial discharge has the capacity to influence ocean properties, marine productivity and the ecosystem {3.3.3}

[START BOX 3.1 HERE]

[3] FOOTNOTE: Some aspects specific to high-mountain regions in northern Russia, Iceland and Alaska are covered in Chapter 2.

Box 3.1: Trends in the Polar Regions' Climate Systems and their Consequences for Atmospheric Links to Lower Latitudes

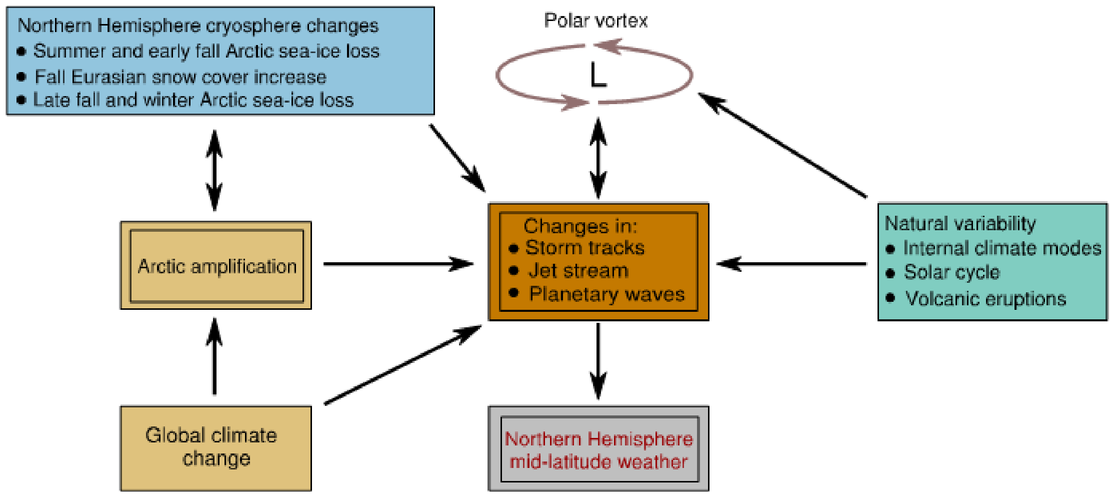
Whilst SROCC is focussed on the ocean and cryosphere, understanding and attributing the observed and predicted changes in the polar regions requires knowledge of drivers in the atmosphere also. Further, understanding the effects that the polar regions have on lower latitudes requires knowledge of the mechanisms by which polar changes are transmitted equatorward. Whilst other parts of this chapter focus on various ocean and cryosphere changes and their consequences; this box assesses the changes of climate systems of the polar regions, including also atmospheric feedbacks and linkages.

For the last two decades, Arctic surface air temperature change has been double the global average, interpreted as a clear indicator of climate change (Notz and Stroeve, 2016; Richter-Menge et al., 2017). This ratio of 2:1 is consistent for projections of the near future in at least the last two generations of global climate assessments (Kattsov and Pavlova, 2015). Stabilizing global temperature rise near 2˚C could slow but not halt further changes in the Arctic climate system (AMAP, 2017). Mechanisms for such Arctic amplification include: reduced summer albedo due to sea ice and snow cover loss, the increase of total water vapour content in the Arctic atmosphere, and a potential change of total cloudiness in summer and increase in the additional heat generated by newly sea-ice free ocean areas that are maintained later into the autumn (Appendix 3.A.1.2). Northward transports of heat and moisture also contribute, as does the lower rate of heat loss to space from the Arctic relative to the sub-tropics (Serreze and Barry, 2011; Pithan and Mauritsen, 2014; Goosse et al., 2018) (see also Appendix 3.A.1.1).

A number of recent events in the Arctic indicate state changes of the Arctic climate system. Annual Arctic temperatures for the past five years (2014-2018) all exceed previous records (Overland and Wang, 2018). Winter (January-March) near-surface temperature anomalies of +6°C (relative to 1981-2010) were recorded in the central Arctic for both 2016 and 2018, nearly double the previous record (Overland and Wang, 2018). These were caused by a split of the tropospheric vortex into two cells making the Arctic more susceptible to influences from subarctic storms (Overland and Wang, 2016). The resulting advection of temperature and moisture from the Pacific and Atlantic Oceans into the central Arctic, which increases downward longwave radiation, delayed sea ice freeze-up and resulted in an unprecedented absence of sea ice. Delayed freeze-up of sea ice in subarctic seas (Chukchi, Barents and Kara) acts as a positive feedback allowing warmer temperatures to progress further toward the North Pole (Kim et al., 2017). Not only were there low sea ice extents in summers since 2007, but now Arctic winter sea ice maximums for the last 5 years are all less than all previous years (Overland, 2018). Multi-year, large magnitude extreme positive Arctic temperatures and sea ice minimums since AR5 provide *high agreement* and *medium evidence* of contemporary states well outside the envelope of previous experience (1900-2017; AMAP (2017)) (see also Section 3.2.1.1).

In contrast to the Arctic, the Antarctic continent has seen less uniform temperature changes over recent decades, with warming over Western Antarctica and the Antarctic Peninsula and weak cooling over East Antarctica (Nicolas and Bromwich, 2014; Jones et al., 2016), though there is *low confidence* in these changes given the sparse in situ records and large internal variability. There is *medium confidence* through a growing body of literature that variability of tropical Pacific sea surface temperatures can influence Antarctic temperature changes (Turner et al., 2016; Smith and Polvani, 2017; Clem et al., 2017a) as well as Antarctic ice shelf and glaciers (Dutrieux et al., 2014; Paolo et al., 2018), the Southern Hemisphere mid-latitude circulation (Schneider et al., 2015a; Raphael et al., 2016; Turney et al., 2017; Clem et al., 2017a; Evtushevsky et al., 2018) and Antarctic sea-ice extent on year-to-year (Stuecker et al., 2017; Turner et al., 2017b; Schneider and Deser, 2018) and decade-to-decade timescales (Meehl et al., 2016; Purich et al., 2016b).

The Southern Annular Mode, Pacific South American mode (by which tropical Pacific convective heating signals are transmitted to high southern latitudes) and zonal-wave 3 are the dominant large-scale atmospheric circulation drivers of Antarctic surface climate and sea-ice changes (Appendix 3.A.1.3). Consistent with AR5, it is *likely* that Antarctic ozone depletion has been the dominant driver of the positive trend in the Southern Annular Mode during austral summer from the late 1970s to the late 1990s (Waugh et al., 2015; Schneider et al., 2015a), with new research suggesting a stronger role of tropical sea surface temperatures since 2000 (Schneider et al., 2015a).

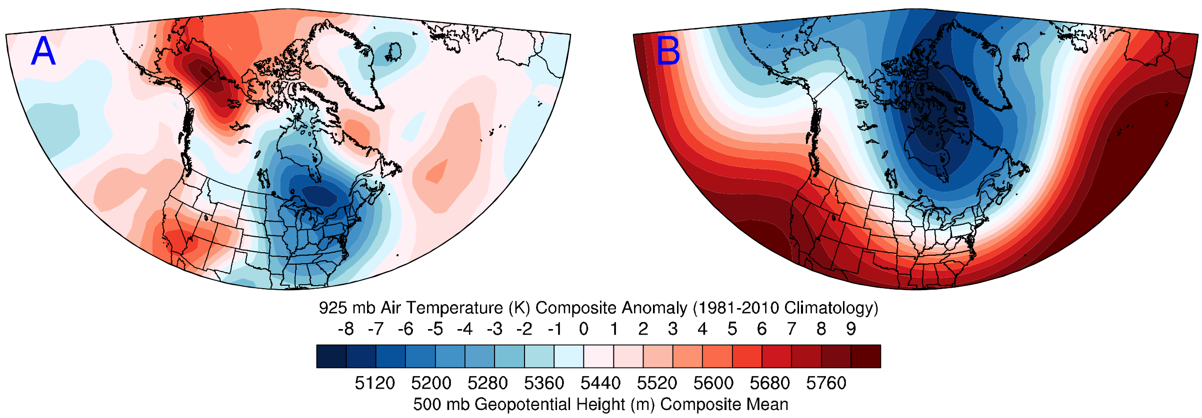


Box 3.1, Figure 1: Complexity of the pathways by which Arctic processes can influence mid-latitude weather. The pathway highlighted with double boxes represents the effect of Arctic amplification directly (by changing the meridional temperature gradient) and/or indirectly (through feedbacks with changes in the cryosphere) on tropospheric wave activity, storm activity and the jet stream at mid- and high latitudes. Two other causes of changes in these that do not involve Arctic amplification are also represented, namely natural variability and the direct influence of global climate change (i.e., including influences outside the Arctic) on general circulation. Bidirectional arrows denote feedbacks (positive or negative) between adjacent elements. Stratospheric polar vortex is represented by 'L' with anticlockwise flow. (Figure adapted from Cohen et al. (2014)).

Potential for Polar Regions and Mid-Latitude Weather Linkages

Since AR5 understanding of Arctic and mid-latitude weather connections has become a societally important topic potentially impacting tens of millions of people (Jung et al., 2015) (Box 3.1, Figure 1), but the meteorological processes involved are complex and full understanding has not yet been developed. Assessments continue to be controversial (National Research Council, 2014; Barnes and Polvani, 2015; Francis, 2017). Arctic forcing from sea ice and snow loss and rising temperatures is clearly increasing, but the link to mid-latitude consequences is mediated by jet stream dynamics; connectivity is reduced by the influence of chaotic internal natural variability and other tropical and oceanic forcing. The potential for Arctic/mid-latitude weather linkages varies for different jet stream patterns (Grotjahn et al., 2016; Messori et al., 2016; Overland and Wang, 2018). Part of the scientific controversy is due to irregular connections in the linkage pathway within and between years.

Considerable literature exists on the potential for cold episodes in eastern Asia from sea ice loss in the Kara Sea (Kim et al., 2014; Kretschmer et al., 2016). There is some analysis of cases between change in the Chukchi Sea and west of Greenland, and cold events in eastern North America (Kug et al., 2015; Ballinger et al., 2018; Overland and Wang, 2018). Such connections seem to be episodic (Cohen et al., 2018) as climatological studies do not show increases in the number of cold events in data or model projections (Ayarzaguena and Screen, 2016; Trenary et al., 2016). A potential North American example was December 2017. Warm temperatures over Alaska and record lack of sea ice (Box 3.1, Figure 2A) helped to anchor the long wave geopotential height pattern (Box 3.1, Figure 2B), which in turn fed cold temperatures into the eastern USA.



Box 3.1, Figure 2: A) 925 mb air temperature anomalies from 8 December 2017 to 7 January 2018. B) matching 500 mb geopotential height pattern

Only a few studies have focused on the potential impact of Antarctic sea-ice changes on the mid-latitude circulation (Kidston et al., 2011; Raphael et al., 2011; Bader et al., 2013; Smith et al., 2017b; England et al., 2018); these find that any impacts on the jet stream are strongly dependent on the season and model examined. England et al. (2018) suggest that the response of the jet stream to future Antarctic sea ice loss may in fact be less seasonal than the response to Arctic sea ice loss.

[END BOX 3.1 HERE]

3.2 Polar Oceans and Sea Ice: Changes, Consequences and Impacts

3.2.1 Observed Changes in Ocean and Sea Ice

3.2.1.1 Sea Ice

Sea ice reflects a high proportion of incident solar radiation back to space, insulates the ocean from the atmosphere, influences thermohaline circulation, limits access to the polar regions, provides habitat for ice-associated species, and is of high importance to the traditional lifestyle of northern residents. The characteristics of sea ice cover differs between the Arctic and Antarctic. The central Arctic Ocean is surrounded by land, and ice circulates within this basin, forced largely by the atmosphere. Conversely, the Antarctic continent is surrounded by sea ice, which interacts with the upper layers of the Southern Ocean and adjacent ice shelves. Differing physical processes and seasonal regimes of Arctic versus Antarctic sea ice result in different observed trends between the two polar regions. Understanding of sea ice variability, changes, and impacts includes rich and pervasive Indigenous Knowledge and Local Knowledge from communities across the circumpolar Arctic {see Cross-Chapter Box 3 in Chapter 1}.

3.2.1.1.1 Extent and concentration

The pan-Arctic loss of sea ice cover is a prominent indicator of climate change (Figure 3.2). Nearly four decades of continuous satellite observations have documented pronounced declines in sea ice extent (the total area of the Arctic with at least 15% sea ice concentration) for each month of the year (Meier et al., 2014; Serreze and Stroeve, 2015; Stroeve and Notz, 2015; Barber et al., 2017) (Figure 3.3) *(very high confidence).* Changes are largest in summer and smallest in winter, with September trends (month with the lowest sea ice cover; 1979 to 2017) of –83,000 $km^2\ yr^{-1}$ (-13.0% per decade relative to 1981-2010 mean), and –41,000 $km^2\ yr^{-1}$ (–2.7% per decade relative to 1981-2010 mean) for March (month with the greatest sea ice cover; Onarheim et al. (2018)). Regionally, summer ice loss is dominated by reductions in the East Siberian Sea (explains 22% of the September trend), and large declines in the Beaufort, Chukchi, Laptev and Kara seas (Onarheim and Årthun, 2017). Winter ice loss is dominated by reductions within the Barents Sea, responsible for 27% of the pan-Arctic March sea ice trends (Onarheim and Årthun, 2017). Reconstructions of the sea ice cover back to 1850 using earlier satellite observations, ship and aircraft observations, ice charts, and whaling records shows that Arctic ice loss over the past 2 decades is *likely* unprecedented in at least 150 years (Walsh et al., 2017).

Figure 3.2: Schematic of some of the major Arctic changes assessed in this section.
(1) strengthening of the circulation of the Beaufort Gyre (Section 3.3.1.3.1);
(2) increasing discharge of freshwater from rivers to the Arctic Ocean (Section 3.3.1.2.2);
(3) strengthening efflux to lower latitudes through Fram Strait (Section 3.3.1.3.1);
(4) increasing glacial loss from Greenland (Section 3.2.1.3);
(5) retreat of sea ice (Section 3.3.1.1.1);
(6) retreat of seasonal snow cover on land (Section 3.4.1.1.1);
(7) changing ice-albedo feedback (Appendix 3.A.1.2);
(8) strengthening transport within the Transpolar Drift (Section 3.3.1.1.4);
(9) increasing oceanic heat transport from North Atlantic (Section 3.3.1.2.1);
(10) increasing oceanic heat transport from North Pacific (Section 3.3.1.2.1);
(11) heating of surface layers via insolation (Section 3.3.1.2.1);
(12) carbon drawdown from atmosphere (Section 3.3.1.2.4);
(13) increasing primary production associated with areas of ice retreat (Section 3.4.4.1.1).

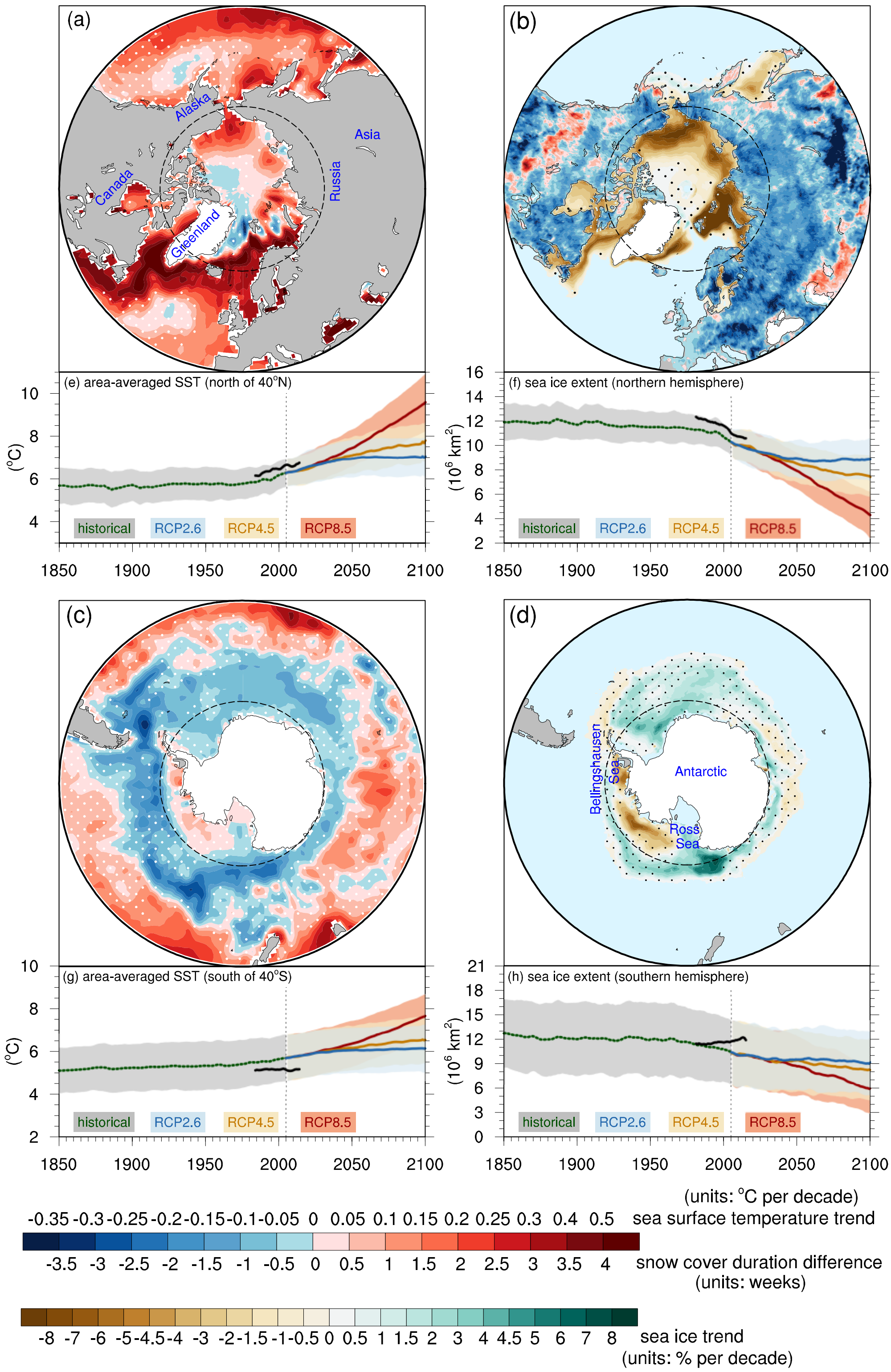


Figure 3.3: (a) Linear trends of Arctic annual-mean sea surface temperature for 1982−2016, in ℃ per decade. (b) Linear trends of Arctic annual-mean sea ice concentration for 1982−2016, in % per decade, alongside the difference in climatological snow cover duration (in weeks) between the period 2006 to 2015 and 1981 to 1990. (c) Linear trends of Southern Ocean annual-mean sea surface temperature for 1982−2016, in ℃ per decade. (d) Linear trends of Antarctic annual-mean sea ice concentration for 1982−2016, in % per decade. Panels (e-h) show the comparable 5-year running mean time series of annual-mean sea surface temperature (area-averaged north of 40°N/south of 40°S) or annual-mean sea ice extent in the northern/southern hemisphere. Black, green, blue, orange, and red curves in (e-h) indicate observations, CMIP5 historical simulation, RCP2.6, RCP4.5, and RCP8.5 projections; shading indicates +/- standard deviation of multi-models. (a) and (c) are from the NOAA Optimum Interpolation Sea Surface Temperature dataset (version 2; Reynolds et al. (2002); https://www.ncdc.noaa.gov/oisst). (b) and (d) are from the NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, Version 3 (https://nsidc.org/data/g02202). Snow cover duration in (b) was derived from a blend of four independent datasets, each covering the 1981-2015 period (Brown et

al., 2003; Takala et al., 2011; Brun et al., 2012; Reichle et al., 2017). The models from CMIP5 include bcc-csm1-1, bcc-csm1-1-m, CanESM2, CCSM4, CESM1-CAM5, CNRM-CM5, GISS-E2-H, IPSL-CM5A-LR, IPSL-CM5A-MR, MIROC5, MIROC-ESM, MIROC-ESM-CHEM, MPI-ESM-LR, MPI-ESM-MR, MRI-CGCM3, NorESM1-M, NorESM1-ME, selected since they provide the relevant variables for analysis and were integrated from 1850.

Approximately 50 to 60% of the observed sea ice loss is driven by increased concentrations of atmospheric greenhouse gases, with the remainder due to natural climate variability (Kay et al., 2011; Notz and Marotzke, 2012; Stroeve et al., 2012b; Stroeve and Notz, 2015; Notz and Stroeve, 2016) (*high confidence*). Anomalous sea ice minima in September are preceded by a wide range of summer atmospheric circulation patterns (Serreze et al., 2016; Ding et al., 2017), with the ice-albedo feedback a key driver in the evolution of summer sea ice cover. Warm and moist air advection (Mortin et al., 2016) and increased downwelling longwave radiation associated with heightened cloudiness and humidity (Kapsch et al., 2013; Hegyi and Deng, 2016) drives the formation of melt ponds (Perovich and Polashenski, 2012) and open water areas (Serreze et al., 2016). This enhances the ice-albedo feedback, leading to more ice melt in summer (Stroeve et al., 2014a) and thinner ice. Once air temperatures drop below freezing in autumn, thermodynamic ice growth for thin ice is enhanced, and later ice freeze-up in autumn delays snowfall accumulation on sea ice leading to a thinner snowpack. These two negative feedbacks help to mitigate sudden and irreversible loss of the Arctic sea ice (Stroeve and Notz, 2015), although winter season ice growth remains sensitive to episodic moisture intrusions (Hegyi and Taylor, 2018).

Antarctic sea ice extent increased between 1979 and 2017 at an annual-mean rate of $20.2 \pm 4.0 \times 10^3$ km^2 yr^{-1} (Comiso et al., 2017), but with strong negative departures in 2016 and 2017 (Turner et al., 2017b) (*very high confidence*). The overall increase is composed of near-compensating regional changes, with rapid ice loss in the Amundsen and Bellingshausen seas outweighed by rapid ice gain in the Weddell and Ross seas (Holland, 2014) (Figure 3.3). The regional trends are strongly seasonal in character (Holland, 2014); only the western Ross Sea has a trend that is statistically significant in all seasons, relative to the variance during the period of satellite observations. Coupled climate models indicate that anthropogenic warming at the surface is delayed by the Southern Ocean circulation, which transports heat downwards into the deep ocean (Armour et al., 2016). This overturning circulation (see Cross-Chapter Box 5), along with differing cloud and lapse rate feedbacks (Goosse et al., 2018), may explain the weak response of Antarctic sea ice cover to increased atmospheric greenhouse gas concentrations compared to the Arctic (*medium confidence*).

The regional pattern of observed Antarctic sea ice trends (Figure 3.3) is closely related to meridional wind trends (*high confidence*) (Holland and Kwok, 2012; Haumann et al., 2014): poleward wind trends in the Bellingshausen Sea keep sea ice close to the coast (Holland and Kwok, 2012) and advect warm air to the sea ice zone (Kusahara et al., 2017), the reverse being true in the Ross Sea. The large sea ice increase in the western Ross Sea is driven by wind trends in west Antarctica, linked to tropical Pacific variability (Simpkins et al., 2014; Coggins and McDonald, 2015; Meehl et al., 2016; Purich et al., 2016b) and the Southern Annular Mode (SAM) (Matear et al., 2015). While the magnitude of these feedbacks is sufficient to explain the expansion of Ross sea ice (Lecomte et al., 2017), there is no evidence of a trigger, so it is not possible to assess whether this is the actual cause (or one of the causes) of the observed trends. Strengthening circumpolar westerly winds and the Southern Annual Mode are linked to ozone depletion over the Southern Hemisphere (Gillett et al., 2013; Christidis and Stott, 2015), and have the potential to affect zonal-mean Antarctic sea ice on two time scales (Ferreira et al., 2015): an initial sea ice expansion followed by a delayed sea ice decrease (*medium confidence*). The longevity of the initial ice expansion phase is highly uncertain, as is the magnitude of its effect (Holland et al., 2017). Ozone depletion may also affect meridional winds (Fogt and Zbacnik, 2014; England et al., 2016), but there is *low confidence* that this explains observed sea ice trends (Landrum et al., 2017). Ocean-sea ice feedbacks may also prolong sea ice anomalies (Goosse and Zunz, 2014).

Proxy reconstructions (Abram et al., 2013; Murphy et al., 2014), ship records (de la Mare, 2009; Edinburgh and Day, 2016) and early satellite images (Gallaher et al., 2014) indicate that there has been a decrease in overall Antarctic sea ice cover since the early 1960s (*medium confidence*), but in the context of the changes during the satellite record, this decline is too modest to be separated from natural variability (Hobbs et al., 2016a) (*high confidence*).

3.2.1.1.2 Age and thickness

First-year ice now occupies up to 60–70% of the Arctic Basin, compared to only 40% in the early and mid-1980s (Stroeve and Dirk, 2018) (*very high confidence*). Since 1979, the proportion of ice at least 5 years old declined from 30% to less than 5% (Maslanik et al., 2011; Stroeve et al., 2012a).

There is *very high confidence* that the ice has thinned because of the shift to younger (and hence thinner) ice, and consistent volume reductions in satellite altimeter measurements (Kwok et al., 2009; Laxon et al., 2013), ocean–sea ice reanalyses (Chevallier et al., 2017) and in situ measurements (Renner et al., 2014). Data from multiple satellite altimeter missions show declines in Arctic Basin ice thickness from 2000 to 2012 of –0.58 ± 0.07 m per decade (Lindsay and Schweiger, 2015). Integration of data from submarines, moorings, and earlier satellite radar altimeter missions shows ice thickness declined across the central Arctic by 65%, from 3.59 to 1.25 m between 1975 and 2012 (Lindsay and Schweiger, 2015). New estimates of ice thickness are available for the marginal seas (up to a maximum thickness of ~1 metre) from low-frequency satellite passive microwave measurements (Kaleschke et al., 2016; Ricker et al., 2017) but data are only available since 2010. The change to thinner, younger sea ice in the Arctic is important because seasonal ice is vulnerable to fragmentation from the passage of intense Arctic cyclones in summer (Zhang et al., 2013) and increased ocean swell conditions which result from increased open water (Thomson and Rogers, 2014). Winter sea ice thickness is also a predictor of summer sea ice extent (Guemas et al., 2014).

In-situ observations of Antarctic sea ice thickness are extremely sparse (Worby et al., 2008). There are no consistent long-term observations from which trends in ice volume may be derived. Calibrated model simulations suggest that ice thickness trends closely follow those of ice concentration (Massonnet et al., 2013; Holland et al., 2014) (*low confidence*).

3.2.1.1.3 Seasonality

There is *high confidence* Arctic sea ice melt season has extended by more than 10.0 days per decade since 1979, largely as a result of later freeze-up (+7.5 days per decade) (Stroeve et al., 2012a). While melt onset trends are generally smaller, they play a large role in the earlier development of open water (Stroeve et al., 2012a; Serreze et al., 2016), and melt pond development (Perovich and Polashenski, 2012), enhancing the ice-albedo feedback (Stroeve et al., 2014a; Liu et al., 2015). Observed reductions in the length of seasonal sea ice cover are reflected in community-based observations of decreased length of time in which activities can safely take place on sea ice (Laidler et al., 2010; Eisner et al., 2013; Fall et al., 2013; Ignatowski and Rosales, 2013).

Changes in the duration of Antarctic sea ice cover largely follow the spatial pattern of sea ice concentration trends (*high confidence*). Ice cover duration in the Amundsen/Bellingshausen Sea region reduced by 3.1 ± 1 days per year from 1979–2011, owing to earlier retreat and later advance; duration in the western Ross Sea increased by 2.5 ± 0.4 days per year, again due to changes in the timing of both advance and retreat (Stammerjohn et al., 2012).

3.2.1.1.4 Motion

Winds associated with the climatological Arctic sea level pressure pattern drive the Beaufort Gyre and the Transpolar Drift Stream, which sequester sea ice within the central Arctic Basin and export sea ice out of the Fram Strait, respectively, with inter-annual variability in atmospheric circulation strongly influencing ice export (Smedsrud et al., 2011; Smedsrud et al., 2017) (*high confidence*). Sea ice drift speeds have increased, both within the Arctic Basin and through Fram Strait (Rampal et al., 2009; Vihma et al., 2012; Olason and Notz, 2014), attributed to thinner ice, reduced ice concentration, and changes in wind forcing (Spreen et al., 2011). There is *medium confidence* in Fram Strait sea ice export estimates of 600,000 to 1 million km^2 of ice annually (approximately 10% of the ice within the Arctic Basin (Smedsrud et al., 2017) because of different trends reported from different datasets over non-standard time periods (Kwok et al., 2013; Krumpen et al., 2016; Smedsrud et al., 2017).

Satellite estimates of sea ice drift velocity show significant trends in Antarctic ice drift (Holland and Kwok, 2012). Increased northward drift in the Ross Sea and decreased northward drift in the Bellingshausen and Weddell seas agree with the respective ice extent gains and losses in these regions, but there is only *medium confidence* in these trends due to a small number of ice drift data products derived from a temporally inconsistent satellite record (Haumann et al., 2016).

3.2.1.1.5 Landfast ice

Immobile sea ice anchored to land or ice shelves is referred to as 'landfast'. Very few long-term records of Arctic landfast sea ice thickness exist, but all exhibit thinning trends in springtime maximum sea ice thickness (*high confidence*). Since the mid-1960s, reported declines are 11 cm per decade in the Barents Sea (Gerland et al., 2008), 3.3 cm per decade along the Siberian Coast (Polyakov et al., 2010), and 3.5 cm per decade in the Canadian Arctic Archipelago (Howell et al., 2016). Over a shorter 1976 to 2007 period, winter season landfast sea ice extent from measurements across the Arctic significantly decreased at a rate of 7% per decade, with the largest decreases in the regions of Svalbard (24% per decade) and the northern coast of the Canadian Arctic Archipelago (20% per decade) (Yu et al., 2013). Svalbard and the Chukchi Sea regions are experiencing the largest declines in landfast sea ice duration (~1 week per decade) since the 1970s (Yu et al., 2013; Mahoney et al., 2014). While most Arctic landfast sea ice melts completely each summer, perennial landfast ice (also termed an 'ice-plug') occurs in Nansen Sound and the Sverdrup Channel in the Canadian Arctic Archipelago. These ice-plugs were in place continuously from the advent of observations in the early 1960s, until they disappeared during the anomalously warm summer of 1998, and they have rarely re-formed since 2005 (Pope et al., 2017). The loss of this perennial sea ice is associated with reduced landfast ice duration in the northern Canadian Arctic Archipelago (Galley et al., 2012; Yu et al., 2013) and increased inflow of multi-year ice from the Arctic Ocean into the northern Canadian Arctic Archipelago (Howell et al., 2013).

Changes in Arctic landfast sea ice have implications for northern residents due to the importance as a platform for travel, hunting, and access to offshore regions (see Section 3.3.5.5). Reports of thinning, less stable, and less predictable landfast ice have been documented by residents of coastal communities in Alaska (Eisner et al., 2013; Fall et al., 2013; Huntington et al., 2017), the Canadian Arctic (Laidler et al., 2010), and Chukotka (Inuit Circumpolar Council, 2014). The impact of changing prevailing wind forcing on local ice conditions has been specifically noted (Rosales and Chapman, 2015) including impacts on the landfast ice edge and polynyas (Gearheard et al., 2013).

Long term records of Antarctic landfast ice are limited in space and time (Stammerjohn and Maksym, 2017), with a high degree of regional variability in reported trends (Fraser et al., 2011) (*low confidence*).

3.2.1.1.6 Snow on ice

Snow accumulation on sea ice inhibits sea ice melt through a high albedo, but the insulating properties limit sea ice growth (Sturm and Massom, 2016) and inhibits photosynthetic light (important for under-ice biota) reaching the bottom of the ice (Mundy et al., 2007). If snow on first-year ice is sufficiently thick, it can depress the ice below the sea level surface, which forms snow-ice due to surface flooding. This process is widespread in the Antarctic (Maksym and Markus, 2008) and the Atlantic sector of the Arctic (Merkouriadi et al., 2017), and may become more common in the Arctic as the ice regime shifts to thinner seasonal ice (Granskog et al., 2018) (*medium confidence*).

Despite the importance of snow on sea ice, surface or satellite-derived observations of snowfall over sea ice, and snow depth on sea ice are lacking (Webster et al., 2014). This gap is the primary source of uncertainty in satellite altimetry-based retrieval of sea ice thickness (Ricker et al., 2015). The primary source of Arctic snow depth on sea ice information are climatologies based on observations collected decades ago (Warren et al., 1999). These are now of limited value due to the rapid loss of multiyear ice across the central Arctic (Stroeve and Dirk, 2018), and large interannual variability in snow depth on sea ice (Webster et al., 2014). Airborne radar retrievals of snow depth on sea ice provide more recent estimates, but spatial and temporal sampling is highly discontinuous (Kurtz and Farrell, 2011). Multi-source time series provide evidence of declining snow depth on Arctic sea ice (Webster et al., 2014) (*medium confidence*) but surface measurements for validation are extremely limited and suggest a high degree of regional variability (Haas et al., 2017; Rösel et al., 2018).

Although there are regional estimates of snow depth on Antarctic sea ice from satellite (Kern and Ozsoy-Çiçek, 2016), airborne remote sensing (Kwok and Maksym, 2014), in situ field measurements (Massom et al., 2001) and ship-based observations (Worby et al., 2008), data are not sufficiently extensive in time nor space to assess changes in snow accumulation on Antarctic sea ice.

3.2.1.2 Ocean Properties

The Polar Oceans are amongst the most rapidly-changing oceans of the world, with consequences for global-scale storage and cycling of heat, carbon and other climatically- and ecologically-important properties (Appendix 3.A.2.1; Appendix 3.A, Figure 2).

3.2.1.2.1 Temperature

AR5 (their section 3.2.2) reported that Arctic surface waters warmed from 1993 to 2007, and observations over 1950–2010 show the Arctic Ocean water of Atlantic origin (i.e., the Atlantic Water Layer) warming starting in the 1970s. Warming trends have continued: August linear trends for 1982–2017 reveal summer mixed-layer temperatures increasing at about 0.5°C per decade (*high confidence)* over large sectors of the Arctic basin that are ice-free in summer (Timmermans et al., 2017) (see also Figure 3.3). This is primarily the result of increased solar warming that accompanies sea-ice loss (Perovich, 2016). Between 1979 and 2011, the decrease in Arctic Ocean albedo corresponded to 6.4 ± 0.9 W m^{-2} more solar energy input to the ocean (*virtually certain*) (Pistone et al., 2014). This excess solar heat *likely* reduces the growth of sea ice by up to 25% in both the Eurasian and Canadian basins (Timmermans, 2015; Ivanov et al., 2016) (see also section 3.2.1.1).

While Atlantic Water Layer temperatures have stabilized since 2008, the total heat content in this layer continues to increase (*medium confidence*), associated with increased volume inflows (Polyakov et al., 2017). Recent changes have been referred to as the 'Atlantification' of the Eurasian Basin; changes are characterized by weaker stratification and enhanced Atlantic Water Layer heat fluxes further northeast. Polyakov et al. (2017) estimate 2 to 4 times larger heat fluxes in 2014-2015 compared with 2007-2008 (*medium confidence*). In the Canadian Basin, the maximum temperature of the Pacific Water Layer increased by about 0.5°C between 2009 and 2013 (Timmermans et al., 2014), with a doubling in integrated heat content over 1987-2017 (Timmermans et al., 2018) (*medium confidence*). Over 2001-2014, heat transport associated with Bering Strait inflow increased by 60%, due to increases in both volume flux and temperature (Woodgate et al., 2015; Woodgate, 2017) (*medium confidence*).

Observations show that during 2006–2013, the Southern Ocean accounted for 67–98% of total heat gain in the upper 2000 m of the global ocean (Roemmich et al., 2015) (*high confidence),* matching estimates from coupled climate models (Frölicher et al., 2015). Southern Ocean warming has been attributed to anthropogenic factors, especially the role of greenhouse gases but also ozone depletion (Swart et al., 2018). Warming is strongest in the upper 2000 m, and peaks in the latitude range 40°S–50°S (Armour et al., 2016) (see Appendix 3.A.2.1; Appendix 3.A, Figures 2 and 3). This contrasts with the surface waters south of the core of the Antarctic Circumpolar Current (ACC), which have warmed on average only by 0.02°C per decade, relative to a global sea surface temperature trend of 0.08°C per decade since 1950 (Armour et al., 2016) (*high confidence),* and which have exhibited cooling in more recent decades (see also Figure 3.3).

There is *high confidence* that the observed pattern of Southern Ocean warming is driven by the upper-ocean overturning circulation and mixing (see Cross-Chapter Box 5), whereby heat uptake at the surface by newly-upwelled waters is transmitted to the ocean interior in intermediate depth layers (Armour et al., 2016). Whilst temperature trends in the ACC itself are driven predominantly by air-sea flux changes (Swart et al., 2018), the warming on its northern side appears too deep to be caused directly by air-sea fluxes (Gille, 2014); instead, heave (vertical movement of density surfaces) is more important (*medium confidence)* (Desbruyeres et al., 2017; Gao et al., 2018). Below the surface south of the ACC, warming extends close to the Antarctic continent, particularly on the shelf along the Amundsen-Bellingshausen Sea where increases of 0.03°C yr^{-1} have been observed between 1975 and 2012 (Schmidtko et al., 2014) (see also Section 3.2.2.3) (*medium confidence*).

The deep ocean below 2000 m globally stores around 19% of the excess anthropogenic heat in the Earth system, with a large part of this (6% of global total heat excess) located in the deep Southern Ocean south of 30°S (Frölicher et al., 2015; Talley et al., 2016) (*medium confidence)*. The AR5-quantified warming of these waters was recently updated (Desbruyeres et al., 2017) to an equivalent heat uptake of 0.07 ± 0.06 W m^{-2} below 2000 m since the beginning of the century, resulting in an extra 34 ± 14 TW south of 30°S from 1980–2012 (Purkey and Johnson, 2013) *(medium confidence)*. The movement of density surfaces has been identified as the main factor influencing Antarctic Bottom Water properties away from Antarctica, while the loss of Antarctic Bottom Water in the Indian and Pacific basins close to the Antarctic continent is consistent with a warming and freshening of these waters (Purkey and Johnson, 2013).

3.2.1.2.2 Salinity

Salinity is the dominant determinant of polar ocean density, and exerts major controls on stratification, circulation and mixing. Changes in salinity are induced by changes in freshwater discharged to the ocean, with the potential to impact water mass formation and circulation (e.g., Thornalley et al. (2018); see also Chapter 6).

Following increases of Arctic Ocean freshwater content reported in AR5 (their Section 3.3.3.3), recent Arctic-wide estimates yield an increase of freshwater (relative to a salinity of 34.8 on the Practical Salinity Scale, used throughout this chapter) of 600 ± 300 km^3 yr^{-1} over 1992 to 2012, with about two-thirds of this trend (*medium confidence*) attributed to a decrease in salinity, and the remainder to a thickening of the freshwater layer (Rabe et al., 2014; Haine et al., 2015; Carmack et al., 2016). The Beaufort Gyre region has seen an increase in freshwater (*medium confidence*) of about 40% (6,600 km^3) over 2003-2017; this and the strengthening of the Gyre have been attributed to strong dominance of clockwise wind patterns over the Canadian Basin between 1997 and 2016 and freshwater accumulation from sea ice melt (Krishfield et al., 2014; Proshutinsky et al., 2015). Freshwater decreases in the East Siberian, Laptev, Chukchi and Kara seas are estimated to be about 180 km^3 between 2003 and 2014 (Armitage et al., 2016) (*low* to *medium confidence*). An increasing trend of 30 ± 20 km^3 yr^{-1} in freshwater flux through Bering Strait, primarily due to increased volume flux, was measured from 1991-2015, with record maximum freshwater influx through Bering Strait in 2014 of around 3,500 km^3 in that year (Woodgate, 2017) (*medium confidence)*. Freshwater fluxes from rivers are also increasing (Section 3.4.1.2.2).

Observed freshening trends in the Southern Ocean are consistent with those reported in AR5; subsequent studies have increased our confidence in their magnitude and sign, and also attributed these changes to anthropogenic influences (Swart et al., 2018). Salinity changes over 1950–2010 show a persistent freshening of surface waters over the whole Southern Ocean, with trends 0.0002–0.0008 yr^{-1} in mode and intermediate waters to below 1500 m (Skliris et al., 2014) (*medium confidence*). Averaged circumpolarly, de Lavergne et al. (2014) observe a freshening south of the ACC of 0.0011 ± 0.0004 yr^{-1} in the upper 100 m since the 1960s (*medium confidence*). This trend intensifies over the Antarctic shelves (except along the western Antarctic peninsula), where freshening up to 0.01 yr^{-1} is observed (Schmidtko et al., 2014). Recently, there has been increased recognition of the importance of sea ice in driving Southern Ocean salinity changes, with Haumann et al. (2016) demonstrating that wind-driven sea ice export has increased by 20 ± 10 Sv from 1982–2008 (where 1 Sv = 10^6 m^3s^{-1}), and that this may have driven freshening of 0.002 ± 0.001 yr^{-1} in the surface and intermediate waters (*medium confidence*). Separately, the central role of sea ice in driving water mass transformations in the Southern Ocean has been highlighted (Abernathey et al., 2016; Pellichero et al., 2018; Swart et al., 2018), hence such changes have the potential to affect overturning circulation (see Cross-Chapter Box 5). Freshwater input to the ocean from the Antarctic Ice Sheet also has the potential to affect the properties and circulation of some Southern Ocean water masses; see section 3.3.3.

3.2.1.2.3 Stratification

Changing stratification in the polar oceans is of key significance to climate and ecosystems. Upper-ocean stratification mediates the transfer of climatically-important properties between the atmosphere and ocean interior, and also is an important factor in determining the rates and distributions of marine primary production.

Arctic Ocean stratification is strongest at the base of the surface mixed layer. General trends between 1979 and 2012 across the entire central Arctic over all seasons, and in the winter in the boundary regions (Chukchi, southern Beaufort and Barents seas) indicate a mixed layer shoaling of about 0.5 to 1 m yr^{-1} (*low* to *medium confidence*), with mixed-layer deepening trends evident in some regions (e.g., the southern Beaufort Sea in summer (Peralta-Ferriz and Woodgate, 2015). Shoaling has been attributed to surface ocean freshening and inhibition of mixed-layer deepening by convection and shear-driven mixing, whilst deepening trends have been attributed to winds that drive offshore transport of surface freshwater (Peralta-Ferriz and Woodgate, 2015). The Atlantification in the Eurasian Basin is associated with weakening stratification in the eastern Eurasian Basin at the top boundary of the Atlantic Water Layer from 2012 to 2016, related to reduced sea-ice cover and increased vertical mixing (Polyakov et al., 2017).

For the Southern Ocean, there is only *limited evidence* for stratification changes in the post-AR5 period. Section 3.3.3 assesses the potential of freshwater discharge from the Antarctic Ice Sheet to influence such stratification.

3.2.1.2.4 Carbon and ocean acidification

Since AR5, new observations have demonstrated the spatial and temporal variability of ocean acidification and controlling mechanisms of carbon systems in different regions (Bellerby et al., 2018). CO_2 dissolves in surface water to form carbonic acid, which, upon dissociation, causes a decrease in pH (acidification) and also the carbonate ion (CO_3^{2-}) concentration, a building block of calcium carbonate ($CaCO_3$, aragonite and calcite as dominant mineral forms) shells and skeletons.

Robbins et al. (2013) showed aragonite undersaturation for about 20% of surface waters in the Canada and Makarov Basins, where substantial sea ice melt occurred. Qi et al. (2017) reported that aragonite undersaturation has expanded northward by at least 5° of latitude, and deepened by about 100 m between the 1990s and 2010. In the East Siberian Arctic Shelf, extreme aragonite undersaturation was observed, reflecting pH changes in excess of those projected in this region for 2100 (Semiletov et al., 2016); this was also observed along the continental margin and traced in the deep Makarov and Canada Basins (Anderson et al., 2017a). Persistent acidification on the East Siberian Arctic Shelf is driven by the degradation of terrestrial organic matter and discharge of Arctic river water with elevated CO_2 concentrations (*high confidence*).

The dissolved inorganic carbon (DIC) concentration increased in the subsurface waters (150-1400m) in the central Arctic between 1991 and 2011 (Ericson et al., 2014) (*high confidence*). The rate of increase was 0.6–0.9 µmol kg^{-1} yr^{-1}in the Arctic Atlantic Water and 0.4–0.6 µmol kg^{-1} yr^{-1} in the upper Polar Deep Water due to anthropogenic CO_2, while no trend was observed in nutrient concentrations in the same water masses. In waters deeper than 2000 m, no significant trend was observed for DIC and nutrient concentrations. Observation-based estimates (MacGilchrist et al., 2014) revealed a net summertime pan-Arctic export of 231 ± 49 Tg C yr^{-1} of DIC across the Arctic Ocean gateways to the North Atlantic; at least 166 ± 60 Tg C yr^{-1} of this was sequestered from the atmosphere (*medium confidence)*.

Since AR5, carbonate system data in annual, seasonal and higher temporal resolution have become available in many Arctic regions, revealing complex processes that influence ocean acidification. Studies have demonstrated highly variable and complex mechanisms including via which sea ice influences carbon cycles including ikaite production and dissolution (Rysgaard et al., 2013; Bates et al., 2014; Geilfus et al., 2016; Fransson et al., 2017). Although the increase of pH and saturation states by biological uptake of CO_2 in the surface water is well documented (Azetsu-Scott et al., 2014; Yamamoto-Kawai et al., 2016) (*high confidence)*, it has been shown that long photoperiods in Arctic summers sustained high pH in kelp forests, slowing ocean acidification (Krause-Jensen et al., 2016).

The major advance in understanding of CO_2 fluxes in the Southern Ocean since AR5 is from the decadal mean estimate (~1 ± 0.5 Pg C yr^{-1}) and linear response to increasing anthropogenic CO_2 prior to 2013 (Takahashi et al., 2012; Lenton et al., 2013) towards new constraints of its seasonal-to-decadal variability (McNeil and Matear, 2013; Landschützer et al., 2014; Landschützer et al., 2015; Ritter et al., 2017; Gregor et al., 2017a; Gregor et al., 2017b). These advances have provided new insight to the earlier model-based assessment of a weakening CO_2 sink in the 1990s (Le Quéré et al., 2007), revealing that it was part of a decadal cycle that reversed in the 2000s (Landschützer et al., 2015; Munro et al., 2015; Williams et al., 2017). (Appendix 3.A.2.2; Appendix 3.A, Figure 4). Resolving the decadal modes of variability has shown that the mean annual flux anomaly of CO_2 in the Southern Ocean can vary from approximately 0.3 ± 0.1 Pg C yr^{-1} in 2001-2002 to -0.4 Pg C yr^{-1} in 2012 (Landschützer et al., 2015). The decadal mode appears to be linked to interannual adjustments in winter maxima possibly linked to the SAM (Landschützer et al., 2015; Gregor et al., 2017a) whilst summer ingassing variability may be linked to adjustments in primary productivity associated with the El Niño/Southern Oscillation (ENSO) (Conrad and Lovenduski, 2015). The decadal variability has the potential to make a significant contribution to explaining the magnitude and timing of the unaccounted carbon determined from the gap between bottom-up and top-down quantifications of the global carbon budget (Le Quéré et al., 2017). An additional driver that has emerged from increasing anthropogenic CO_2 fluxes is changes to the buffering capacity of the Southern Ocean; this has started to increase the amplitude of the seasonal cycle of pCO_2 over the past 3 decades (1.1 ± 0.3 μatm per decade) (McNeil and Sasse, 2016; Landschützer et al., 2018) (Appendix 3.A.2.3). The *confidence* levels for the decadal modes and the trends in decreasing buffering capacity are *medium* to *high,* but data sparseness and model limitations make the *confidence* on potentially important links to seasonal drivers *low* to *medium*.

Observational products are largely based on coordinated gridded ship-based data products (Bakker et al., 2016); significant data gaps in these, especially in the wintertime Southern Ocean, reduce the confidence levels on the contemporary trends and variability (Gruber et al., 2017; Ritter et al., 2017; Fay et al., 2018). Recent initiatives based on biogeochemical-enabled floats suggest that ship-based observations overlooked higher than expected CO_2 outgassing fluxes south of the Polar Front in winter (Williams et al., 2017; Gray et al., 2018). The *confidence* level of this finding is *low/medium* pending independent confirmation.

Recent reassessments of carbon storage in the Southern Ocean reveal strong sensitivity to changes in overturning circulation (Cross-Chapter Box 5), with anthropogenic and natural carbon being highly variable (30–100%) but out of phase on decadal timescales (DeVries et al., 2017; Tanhua et al., 2017); see also Appendix 3.A.2.4). Both mode and intermediate waters are especially influential in this changing storage, also showing a high sensitivity to meridional shifts in the wind stress (Swart et al., 2014; Swart et al., 2015a; Tanhua et al., 2017). Zonal basin differences in the variable uptake and storage of anthropogenic carbon are not well resolved; the presence of subduction hotspots that suggest that basin-wide studies may be underestimating the importance of mode water subduction as a principal storage mechanism has been highlighted (Langlais et al., 2017).

Current estimates of the strengthening impacts of Southern Ocean acidification are illustrated by the $3.9 \pm 1.3\%$ decrease in derived calcification rates (1998 – 2014) (Freeman and Lovenduski, 2015). These changes have strong regional character with decreases in the Indian and Pacific Sectors (7.5–11.6%) and increases in the Atlantic Ocean ($14.3 \pm 5.1\%$). This period coincides with the invigoration of CO_2 uptake by the Southern Ocean (Landschützer et al., 2015; Gregor et al., 2017a) but its regional character highlights that long-term trends are a complex interplay of regional ecological, biogeochemical and physical drivers.

3.2.1.3 Ocean Circulation

The major elements of Southern Ocean circulation on different spatial and temporal scales are assessed in Cross-Chapter Box 5; Arctic Ocean circulation is considered here. Processes occurring in the Arctic, such as the discharge to the ocean of freshwater from the Greenland Ice Sheet, have the potential to impact on the formation of the headwaters of the Atlantic Meridional Overturning Circulation (see Chapter 6), and also can impact on the structure and operation of the marine ecosystem with implications for commercially-harvested species (Sections 3.2.3, 3.2.4).

Satellite data indicate a general strengthening of the surface geostrophic currents in the Arctic basin (Armitage et al., 2017). Between 2003 and 2014, the strength of the Beaufort Gyre circulation approximately doubled, with similar increases in the strength of the southward surface flow at Fram Strait (Armitage et al., 2017) (*medium confidence*). Over 2001-2014, annual Bering Strait volume transport from the Pacific to the Arctic Ocean increased from 0.7×10^6 $m^3 s^{-1}$ to 1.2×10^6 $m^3 s^{-1}$ (Woodgate et al., 2015) (*medium confidence*).

Mesoscale eddies are important components of the ocean system, exerting strong influences on circulation, mixing and the transport of climatically- and ecologically-important tracers. Increased wind power input to the Arctic Ocean system can in principle be compensated by the production of eddy kinetic energy; analysis of observations in the Beaufort Gyre region suggest compensation by eddies is *about as likely as not* (Meneghello et al., 2017). Data of sufficiently high temporal and spatial variability is limited in the boundary regions of the Arctic Ocean, precluding estimates of eddy variability on a basin-wide scale. In the central basin regions, a statistically significant higher concentration of eddies was sampled in the Canadian Basin compared to the Eurasian Basin between 2003 and 2014; further, a medium correspondence was found between eddy activity in the Beaufort Gyre region and intensified gyre flow (Zhao et al., 2014; Zhao et al., 2016).

In contrast to the Southern Ocean (see Cross-Chapter Box 5), there is comparatively little knowledge on changing Arctic frontal positions and current cores since AR5. The notable exception is that the center of the Beaufort Gyre in 2013 was located about 300 km to the northwest of its position in 2003 (*medium confidence)*, contemporaneous with changes in its freshwater accumulation and alterations in wind forcing (Section 3.2.1.2.2) (Armitage et al., 2017).

[START CROSS-CHAPTER BOX 5 HERE]

Cross-Chapter Box 5: Southern Ocean Circulation: Drivers, Changes and Implications [TBC]

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The Southern Ocean is disproportionately important in global climate and ecological systems, being the major connection linking the Atlantic, Pacific and Indian Oceans in the global circulation. The Southern Ocean is relatively important for the uptake of heat and carbon by the ocean, even beyond what would be expected given its vast size (e.g., Frölicher et al., 2015). The circulation in the Southern Ocean is comprised of an eastward flowing mean current characterized by strong small-scale transient features known as eddies (Figure CB5.1). The mean flow circumnavigates Antarctica as a series of sinuous, braided jets that, taken together, form the world's largest ocean current, the Antarctic Circumpolar Current (ACC). The ACC transports approximately $173.3 \pm 8.7 \times 10^6$ m^3 s^{-1} (Donohue et al., 2016) eastward in geostrophic balance with the contrasting properties between the waters around Antarctica and inside the subtropical gyres to the north of ACC, driven in part by a combination of strong westerly winds and the production of dense water near Antarctica.

The Southern Ocean is also the key region globally for the upwelling of interior ocean waters to the surface, enabling waters that were last ventilated in the pre-industrial era to interact with the industrial-era atmosphere and the cryosphere. New water masses are produced that sink back into the ocean interior, with the export of both extremely cold and dense Antarctic Bottom Water and the lighter Antarctic Intermediate Water and Subantarctic Mode Waters (Figure CB5.1) representing important pathways for surface properties to be sequestered from further interactions with the atmosphere for decades to millennia. This upwelling and sinking constitutes a two-limbed overturning circulation that is driven by a combination of winds and buoyancy forcing, and which is the mechanism by which much of the global deep ocean is renewed.

Southern Ocean overturning circulation plays a strong role in mediating climate change via the transfer of heat and carbon (including that of anthropogenic origin) with the atmosphere (Sections 3.2.1.2; 5.2.2.2); it also has an impact on sea ice extent and concentration, with implications for climate via albedo (Section 3.2.1.1). It acts to oxygenate the ocean interior, and supplies unused nutrients that support a significant fraction of primary production in the rest of the world ocean (Section 5.2.2.2). The upwelling waters in the overturning bring heat to the Antarctic shelf seas, with consequences for ice shelves, marine-terminating glaciers and the stability of the Antarctic Ice Sheet {3.3.1}. The lower limb of this overturning circulation supplies Antarctic Bottom Water that forms the abyssal layer of much of the world ocean (Section 3.2.1.2; 5.2.2.2). Both horizontal and overturning circulations in the Southern Ocean exert influence on the structure and function of the marine ecosystem, via determining habitat and controlling connectivity over ranges of spatial scales, and the strong meridional temperature gradients across the ACC have been invoked as a key factor influencing the level of vulnerability of Antarctica to invasive species (Section 3.2.3.2).

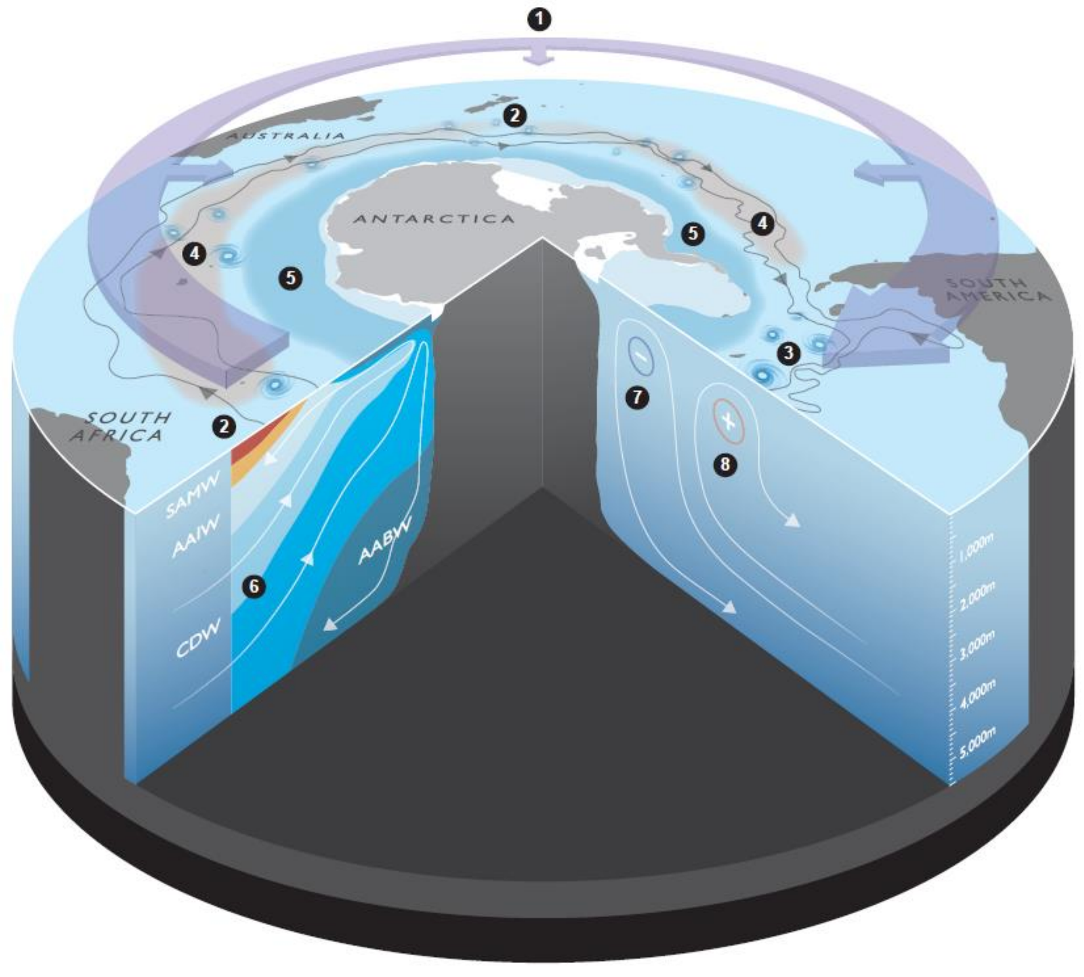

Figure CB5.1. Schematic of some of the major Southern Ocean elements and changes discussed here and in Chapters 3 and 5.
1. Strong circumpolar westerly winds, which have increased and contracted polewards
2. Horizontal circumpolar flow of the Antarctic Circumpolar Current
3. Southern Ocean eddy field, which has intensified in recent decades
4. Warming of surface waters toward northern part of circumpolar Southern Ocean
5. Freshening and delayed warming of surface layers in southern part of circumpolar Southern Ocean
6. Southern Ocean overturning circulation, with upwelling of Circumpolar Deep Water (CDW), and formation and export of Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) in its upper cell, and Antarctic Bottom Water (AABW) in its lower cell
7. Reduction in export of deep and abyssal waters in the lower cell of the overturning circulation
8. Strengthening of the upper cell of the overturning circulation

Trends in the atmospheric forcing of the Southern Ocean are dominated by an increase in westerly winds in recent decades. However, there is no evidence that this enhanced wind stress has altered the ACC transport, the annual mean value of which appears to be remarkably stable in the instrumental period (Chidichimo et al., 2014; Koenig et al., 2014; Donohue et al., 2016) (*medium confidence*). Indeed, there is evidence for only minimal changes in ACC transport since the last glaciation (McCave et al., 2013). Theoretical predictions and high-resolution ocean modelling suggest that the insensitivity of the ACC to changes in wind stress is a consequence of eddy saturation (Munday et al., 2013), wherein the time-mean state of the ocean remains close to a marginal condition for eddy instability and additional energy input from stronger winds cascades rapidly into the smaller-scale eddy field. Satellite measurements of eddy kinetic energy over the last two decades are consistent with this expectation, showing a statistically significant upward trend in eddy energy in the Pacific and Indian Ocean sectors of the Southern Ocean (Hogg et al., 2015) (*medium confidence*). This is supported by eddy-resolving models, which also show a marked regional variability (Patara et al., 2016), and there is evidence that local hotspots in eddy energy, especially downstream of major topographic features, including the Drake Passage, Kerguelen Plateau, Campbell Plateau and the East Pacific Rise, may dominate the regional fields (Thompson and Naveira Garabato, 2014).

It is challenging to measure the Southern Ocean overturning directly, and the upper cell was reported incorrectly in AR5 as having slowed. However, indirect estimates since AR5 provide support for the increase in the upper ocean overturning proposed by Waugh et al. (2013). Waugh (2014) and Ting and Holzer (2017) suggest that over the 1990s-2000 water mass ages changed in a manner consistent with an increase in upwelling and overturning. However, inverse analyses suggest that overturning experiences significant inter-decadal variability in response to wind forcing (DeVries et al., 2017); combined with the indirect nature of observational estimates, such indirect measures give *low confidence* in there having been an acceleration in overturning.

Available evidence indicates that the volume of Antarctic Bottom Water in the global ocean has decreased (Purkey and Johnson, 2013; Desbruyeres et al., 2017) (*medium confidence*), thinning at a rate of 8.1 m yr^{-1} since the 1950s (Azaneu et al., 2013); this reduction in volume has continued in recent years (Figure 5.3). This suggests that the production and export of this water mass has probably slowed, though direct observational evidence is difficult to obtain (*low confidence*). The large-scale impacts of Antarctic Bottom Water changes include a potential modulation to the strength of the Atlantic Meridional Overturning Circulation (e.g., Patara and Böning (2014); see also Section 5.2.2.2.1).

AR5 assessed that there was *medium confidence* that the mean position of the ACC had moved southwards in response to a contraction of the Southern Ocean circumpolar winds. Such movements can in principle have profound effects on marine ecosystems via, e.g., changing habitat ranges for different species (e.g., Cristofari et al. (2018); Section 3.2.3.2; Table 3.1). Since AR5, however, substantial contrary evidence has emerged. A variety of methods applied to satellite data have found no long-term trend and no statistically significant correlation of ACC position with winds (Gille, 2014; Chapman, 2017; Chambers, 2018). The discrepancy between these studies and those assessed in AR5 appears to be caused by issues associated with using a fixed sea surface height contour as a proxy for frontal position in the presence of strongly eddying fields (Chapman, 2014) and large-scale trends in sea surface height due to steric change. These recent findings do not preclude more local changes in frontal position, but this report now reassesses that it is *unlikely* that there has been a statistically significant net southward movement of the mean ACC position over the past 20 years.

Projections of future trends in the Southern Ocean are dominated by the potential for a continued strengthening of the westerly winds (Bracegirdle et al., 2013), as well as a combination of warming and increased freshwater input from both increased net precipitation in the Southern Ocean and net loss of ice from of Antarctica (Downes and Hogg, 2013). Dynamical considerations and numerical simulations indicate that if further increases in the westerly winds are indeed sustained, then it is *very likely* that the eddy field will continue to grow in intensity, with potential consequences for the upper-ocean overturning circulation and transport of tracers (including heat, carbon, oxygen and nutrients), and *likely* that the mean ACC flow will remain insensitive to winds.

The considerable CMIP5 inter-model variations in Southern Ocean time-mean circulation projections reported in AR5 (Meijers et al., 2012; Downes and Hogg, 2013) remain largely unchanged. Some of the differences in projected changes are strongly correlated with model biases in the various models' ability to simulate the historical state of the Southern Ocean (Russell et al., 2018), suggesting that improvements in future generations of coupled models (e.g., CMIP6) should lead to improved confidence in projected changes in the Southern Ocean. CMIP5 models suggest that time-mean subduction and transport of upper Southern Ocean water masses may increase by up to 20% in future (Downes and Hogg, 2013), but model performance is limited by the representation of eddy processes (Gent, 2016; Downes et al., 2018). The formation and export of Antarctic Bottom Water is predicted to continue decreasing (Heuzé et al., 2015) due to warming and freshening of surface source waters near the continent. These are, however, some of the most poorly-represented processes in the simulated global ocean; *low confidence* is therefore ascribed to the CMIP5-based model projections of future Southern Ocean circulation and water mass projections.

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[END CROSS-CHAPTER BOX 5 HERE]

3.2.2 Projected Changes in Ocean and Sea Ice

3.2.2.1 Sea Ice

Historical simulations from CMIP5 models identify declines in total Arctic sea ice extent and thickness (see Section 3.2.1.1.1; 3.2.1.1.2; Figure 3.3) which approximate observations (Massonnet et al., 2012; Stroeve et al., 2012b; Stroeve et al., 2014b; Stroeve and Notz, 2015), but the rates of change, sea ice thickness patterns, general features of Arctic atmospheric circulation, and ice drift rates are not well simulated (Stroeve et al., 2014b). Although aerosols have influenced historical sea ice trends (Gagné et al., 2017), reductions in Arctic sea ice extent scale linearly with both global temperatures and cumulative CO_2 emissions in simulations and observations. However, the observational uncertainty of sea ice sensitivity is quite large (Niederdrenk and Notz, 2018), and the modeled sensitivity (ice loss per unit of warming) is too low in most models (Rosenblum and Eisenman, 2017), due in large part to models underestimating the increase in downwelling longwave radiation associated with increases in atmospheric CO_2 (Notz and Stroeve, 2016).

CMIP5 models project continued declines in Arctic sea ice through the end of the century (Figure 3.3) (Notz and Stroeve, 2016) (*high confidence*). There is a large spread in the timing of when the Arctic may become ice free in the summer (and for how long during the season) in the future (Massonnet et al., 2012; Stroeve et al., 2012b; Overland and Wang, 2013) as a result of internal climate variability (Notz, 2015; Swart et al., 2015b), scenario uncertainty (Stroeve et al., 2012b; Liu et al., 2013), and model uncertainties related to sea ice dynamics (Rampal et al., 2011; Tandon et al., 2018) and thermodynamics (Massonnet et al., 2018). Internal climate variability alone results in an uncertainty of approximately 20 years in the timing of seasonally ice-free conditions (Notz, 2015; Jahn et al., 2016). The clear link between the summer sea ice extent and cumulative CO_2 emissions provide a basis for when ice-free conditions may be expected. After 10 years of stabilized warming at a 2°C global temperature increase, the Arctic is *very likely* to have an ice-free September (Mahlstein and Knutti, 2012; Jahn et al., 2016; Notz and Stroeve, 2016). For stabilized warming at 1.5°C global warming target, sea ice in September is *very likely* to be present at end of century, but individual ice-free years are still projected to occur (Notz and Stroeve, 2016; Sanderson et al., 2017; Jahn, 2018; Sigmond et al., 2018). Model studies show with *high confidence* that a temporary temperature overshoot of a given warming target has no lasting impact on ice cover (Armour et al., 2011; Ridley et al., 2012; Li et al., 2013).

CMIP5 models show a wide range of mean states and trends in Antarctic sea ice (Turner et al., 2012; Shu et al., 2015). Ensemble means across multiple models show a decrease in total Antarctic sea ice extent during the satellite era, in contrast to the observed increase. Interannual sea ice variability in the models is much larger than observations (Zunz et al., 2013), which may mask disparity between models and observations. Internal variability (Polvani and Smith, 2013; Zunz et al., 2013), and model sensitivity to warming (Rosenblum and Eisenman, 2017) are also important sources of uncertainty. Regional trends of Antarctic sea ice are not captured by the models, particularly the decrease in the Bellingshausen Sea and the expansion in the Ross Sea (Hobbs et al., 2015). This is because the regional pattern of trends is linked to random internal climate variability emanating from the tropical Pacific (Schneider and Deser, 2018). There is a very wide spread of model responses in the Weddell Sea (Hobbs et al., 2015; Ivanova et al., 2016), a region with complex ocean-sea ice interactions that many models do not replicate (de Lavergne et al., 2014).

Since AR5, research has focussed on explaining observed and simulated trends in Antarctic sea ice over the historical period, with little new research on projections. There is *low confidence* in projections of Antarctic sea ice because there is no consensus on the drivers of observed changes (see Section 3.2.1.1.), and due to a number of model deficiencies related to stratification (Sallée et al., 2013b), freshening by ice shelf melt water (Bintanja et al., 2015), cloud processes (Schneider and Reusch, 2015b), wind variability, limiting the sea ice response in the regions with the greatest observed trends (Purich et al., 2016a; Purich et al., 2016b; Schroeter et al., 2017). This uncertainty reduces confidence in projections of Antarctic Ice Sheet surface mass balance, because sea ice biases affect Antarctic temperature and precipitation trends (Bracegirdle et al., 2015). They may also impact projected changes in the Southern Hemisphere atmosphere jet (Bracegirdle et al., 2018; England et al., 2018), with implications for the Southern Ocean overturning circulation and the ACC (Cross-Chapter Box 5).

[START BOX 3.2 HERE]

Box 3.2: Polynyas

Arctic Coastal Polynyas

Polynyas (areas of open water surrounded by sea ice) form regularly in many Arctic regions during winter and spring due to a combination of latent (wind) and sensible (heat) effects (Barber and Massom, 2007), and are areas of intense air-ice-ocean exchange (Morales Maqueda et al., 2004). The warm and exposed ocean surface creates very high heat fluxes and sea ice formation rates during winter, releasing brine and creating dense water that helps ventilate the stratified Arctic Ocean (Barber et al., 2012). On the shallow Siberian shelves, the ocean surface layer is dominated by river runoff, thus the mainly wind driven coastal polynyas transport relatively fresh water into the shelf's bottom layer to produce brine-enriched but relatively lower salinity shelf bottom waters (Bauch et al., 2012; Janout et al., 2015). This process maintains the Arctic Ocean halocline (Bauch et al., 2011), which insulates the sea-ice cover from the heat of the underlying Atlantic derived waters.

Polynyas are projected to change in different ways depending on regional ice conditions and processes responsible for formation. They may cease to exist where seasonal sea ice disappears, or evolve to become part of a marginal sea ice zone due to changes in ice dynamics (i.e., the North Water polynya and the Circumpolar Flaw Lead). Further reductions in sea ice are projected for Arctic shelf seas which have lost ice in recent decades (Onarheim et al., 2018). By 2100 under RCP8.5, all of Alaska's northern shore is projected to be ice-free all year, as are the Kara and Barents Seas and Baffin Bay, while the Siberian coast still has approximately six months of sea ice cover (Barnhart et al., 2015). New or enlarged polynyas could result in regions where thinner ice becomes more effectively advected offshore, or where marine terminating glaciers increase land ice fluxes to the marine system.

In spring and summer, polynyas are the first areas exposed to solar insolation. The spring phytoplankton bloom therefore starts earlier (so long as nutrients are available to the euphotic zone and the polynya remains open) and the ocean is well-ventilated and often nutrient rich, so the entire biological range from phytoplankton to seabirds to marine mammals thrive in polynya waters (Stirling, 1997; Arrigo and van Dijken, 2004; Karnovsky et al., 2009). Secondary production and upper food web processes are typically adapted to the early availability of energy to the system with arrival of higher trophic species (Asselin et al., 2011).

Because of the abundance of marine food resources including seals, whales, and fish in and around polynyas, they have been regular areas for hunting by Arctic peoples thousands of years (Barber and Massom, 2007). Recent implementation of Inuit-led marine management areas acknowledge the Inuit knowledge of polynyas, and recognize the potential for development of fisheries and non-renewable resources in polynya systems, provided these activities minimize harm on the environment and wildlife. The Inuit Circumpolar Council's Pikialasorsuaq Commission is an example of a proposal to develop an Inuit management area in the North Water Polynya (see Cross-Chapter Box 3 in Chapter 1).

Antarctic Coastal Polynyas

The Antarctic continent is surrounded by coastal polynyas, which form in the lee of coastal features that protrude into the westward coastal current (Tamura et al., 2008; Nihashi and Ohshima, 2015). Intense ice growth within these polynyas contributes to the production of Antarctic Bottom Water, the densest and most voluminous water mass in the global ocean (Jacobs, 2004; Nicholls et al., 2008; Orsi and Wiederwohl, 2009; Ohshima et al., 2013). Sea ice production is greatest in polynyas of the Ross and Weddell seas and around East Antarctica (Tamura et al., 2008; Drucker et al., 2011; Nihashi and Ohshima, 2015). Ice production in the largest polynya, in the Ross Sea, has increased significantly in recent decades (*high confidence*), driven by increased southerly winds (Drucker et al., 2011; Haumann et al., 2016).

Antarctic coastal polynyas are biological hot-spots that support high rates of primary production (Arrigo and van Dijken, 2003) due to a combination of both high light (Park et al., 2017) and high nutrient levels, especially iron (Alderkamp et al., 2015; Gerringa et al., 2015). Melting ice shelves are the primary supplier of iron to coastal polynyas, more important than either melting sea ice or sediment resuspension via convective mixing in winter (Arrigo and van Dijken, 2015).

As ice shelves retreat, the polynyas created in their wake also increase local primary production. The new polynyas created after the collapse of the Larsen A and B ice shelves are as productive as other Antarctic shelf regions, *likely* increasing organic matter export and altering marine ecosystem evolution (Cape et al., 2013). The recent calving of Mertz Glacier Tongue in East Antarctica has altered sea ice and ocean stratification (Fogwill et al., 2016) such that polynyas there are now twice as productive (Shadwick et al., 2017).

The productivity associated with these polynyas is a critical food source for some of the most abundant top predators in Antarctic waters, including penguins, albatross, and seals (Raymond et al., 2014; Labrousse et al., 2017; Malpress et al., 2017). However, only a fraction of the carbon fixed by phytoplankton in coastal polynyas is consumed by upper trophic levels. The rest sinks to the seafloor where it is remineralized or sequestered (Shadwick et al., 2017), or is advected off the shelf (Lee et al., 2017b). Given the high amount of residual macronutrients in polynya surface waters, future changes in ice shelf melt rates could increase water column productivity (Alderkamp et al., 2015), dramatically influencing Antarctic coastal ecosystems and increasing the ability of continental shelf waters to sequester atmospheric carbon dioxide (Arrigo and van Dijken, 2015).

The Weddell Polynya

The Weddell Polynya is a large area of open water within the winter ice pack of the Weddell Sea (at approximately 60ºS, 15ºW). This is unusual in Antarctica, where most polynyas form along the coast. The polynya opens intermittently, and remained open from 1974 to 1976, with an area of 0.2–0.3 million km^2 (Carsey, 1980). An area of low sea ice concentration appeared in this area following extreme low Antarctic sea ice extent in spring 2016 and 2017, but did not occur in 2018. The polynya forms close to the Maud Rise seamount, and may be caused by ocean eddies creating sea ice divergence over deep ocean water (Holland, 2001). Around Maud Rise, the ocean is weakly stratified, and sea ice formation causes mixing of warmer deep waters at the surface, sufficient to melt newly-formed sea ice (Martinson et al., 1981). Passing winter storm systems may also influence stratification and rapidly ventilate heat, leading to periods of reduced ice cover (Wilson et al., In review). These processes allow the Weddell Polynya to occur in some years, cause deep ocean convection that releases heat from the deep ocean to the atmosphere (Smedsrud, 2005), and may contribute to the uptake of anthropogenic carbon (Bernardello et al., 2014).

In some CMIP5 models, phases of Weddell polynya activity appear for decades or centuries at a time, and then cease for a similar period (Reintges et al., 2017b). The observational era is not sufficiently long to rule out this behaviour. Models indicate that under anthropogenic climate change, surface freshening caused by increased precipitation reduces the occurrence of the Weddell polynya (de Lavergne et al., 2014). There are systematic biases in modelled ocean stratification due to lack of realistic freshwater input from ice shelves and melting icebergs, producing *low confidence* in the future Weddell Polynya projections (Reintges et al., 2017a).

[END BOX 3.2 HERE]

3.2.2.2 Physical Oceanography

Consistent with the projected sea ice decline, there is *high confidence* that the Arctic Ocean will warm significantly towards the end of this century at the surface and in the deeper layers. Most CMIP5 models are able to capture the seasonal changes in surface heat and freshwater fluxes for the present day climate, and show that the excess summer solar heating is used to melt sea ice, in a positive ice-albedo feedback (Ding et al., 2016). Using RCP8.5, Vavrus et al. (2012) found that the Atlantic layer temperature is projected to warm by 2.5°C at around 400 m depth at the end of the century, but only by 0.5°C in the surface mixed layer. Consistent results for lower Atlantic Water layer warming was found by Koenigk and Brodeau (2014) for RCP2.5 (+0.5°C), RCP4.5 (+1.0°C) and RCP8.5 (+2.0°C).

Poleward ocean heat transport contributes to Arctic Ocean warming (*medium confidence*). Comparing 20 CMIP5 models for RCP8.5, Nummelin et al. (2017) found a 2°C –6°C range in Arctic amplification of surface air temperature north of 70°N, consistent and associated with increased ocean heat transport. Comparing 26 different CMIP5 models for RCP4.5, Burgard and Notz (2017) found that ocean heat transport changes explain the Arctic Ocean multi-model mean warming, but that differences between models are compensated by changes in surface fluxes. Increased ocean heat transport into the Barents Sea beyond 2020 is suggested as the main mechanism based on one CMIP5 model (Koenigk and Brodeau, 2014). Based on 4 CMIP5 models, the Barents Sea becomes ice-free during winter beyond 2050 under RCP8.5 (Onarheim and Årthun, 2017), to which the main response is an increased ocean to atmosphere heat flux and related surface warming (Smedsrud et al., 2013). When the winter sea ice disappears the heat loss cannot increase further, and the excess ocean heat continues into the Arctic Basin (Koenigk and Brodeau, 2014). The ocean heat transport will also increase through the other Arctic gateways (Bering Strait, Fram Strait, and the Canadian Archipelago), but the increase appears smaller than in the Barents Sea.

The surface mixed layer of the Arctic Ocean is expected to freshen in the future because an intensified hydrological cycle will increase river runoff (Haine et al., 2015) (*medium confidence*). The related increase in stratification has the potential to contribute to the warming of the deep Atlantic Water layer, as upward vertical mixing will be reduced (Nummelin et al., 2016). There are, however, biases in salinity of ~1 across the Arctic Basin for the present-day climate (Ilicak et al., 2016) in forced global ice-ocean models with comparable configurations to CMIP5, with all models being too saline at the surface in the Canada Basin and too fresh at 50–400 m depth, suggesting limited predictive skill for the Arctic freshwater cycle.

CMIP5 modelling (Figure 3.3) indicates that observed Southern Ocean warming trends will continue under RCP4.5 and RCP8.5, leading to 1°C–3°C warming by 2100 mostly in the upper ocean (Sallée et al., 2013b). Model projections demonstrate a similar distribution of heat storage to historical observations, notably focused in deep pools north of the Subantarctic Front (e.g., Armour et al., 2016). Antarctic Bottom Water becomes coherently warmer by up to 0.3°C by 2100 across the model ensemble under RCP8.5 (Heuzé et al., 2015). The upper ocean water masses also become considerably fresher (salinity decrease of approximately 0.1) (Sallée et al., 2013a) with an overall increase in stratification and shoaling mixed layer depths (Sallée et al., 2013b). Although the sign of model changes appear mostly robust, there is *low confidence* in magnitude due to the large inter-model spread in projections and significant warm biases in historical water mass properties (Sallée et al., 2013b) and sea surface temperature, which may be up to 3°C too high in the historical runs (Wang et al., 2014). Projections of changes in Southern Ocean circulation are discussed in Cross-Chapter Box 5.

3.2.2.3 Carbon and Ocean Acidification

While the large decrease of pH and aragonite saturation in the Arctic were projected using global models in AR5 and the influence of sea ice reduction rate to ocean acidification was demonstrated (Yamamoto et al., 2012), regional models have been developed subsequently. The Canadian Arctic Archipelago and Baffin Bay show greatest rates of acidification and saturation state decline as a result of melting sea ice (Popova et al., 2014) (*high confidence*). In the Canada Basin, projections under RCP8.5 forcing show reductions in the bidecadal mean surface pH from about 8.1 in 1986–2005 to 7.7 by 2066–2085 and aragonite saturation from 1.52 to 0.74 during the same period (Steiner et al., 2014). A shoaling of the aragonite saturation horizon of approximately 1200 m and a large increase in area extent of undersaturated surface waters were projected in the Nordic Sea, with a simulated pH change in the surface water is –0.19 from 2000 to 2065 (Skogen et al., 2014).

CMIP5 models project that the uptake of CO_2 by the Southern Ocean will increase from the contemporary 0.91 Pg C yr^{-1} to 2.38 (1.65–2.55) Pg C yr^{-1} by 2100, but the growth in uptake will stop in about 2070 corresponding to cumulative CO_2 emissions of 1600 GtC (Kessler and Tjiputra, 2016; Wang et al., 2016b).

This halt in the increase in the uptake rate of CO_2 is linked to the feedback from both reduced buffering capacity and increased upwelling rates of Circumpolar Deep Water (Hauck and Volker, 2015) (see also Cross-Chapter Box 5). Contemporary biases in the fluxes of CO_2 in CMIP5 models in the Southern Ocean suggest the *confidence* levels for these projections to be *medium* (Mongwe et al., 2018).

The onset of aragonite undersaturation in the Southern Ocean is influenced by the seasonal cycle of carbonate as well as by reduced buffering capacity on the seasonal cycle linked to anthropogenic CO_2 (Sasse et al., 2015; McNeil and Sasse, 2016). One of the most important additional outcomes from decreasing buffering capacity is an amplification of the seasonal variability of pCO_2 and pH (Hauck and Volker, 2015; McNeil and Sasse, 2016; Landschützer et al., 2018). This amplification accelerates the onset of hypercapnia (i.e., high pCO_2 levels; pCO2 > 1000 µatm) to nearly 2 decades ahead of atmospheric forcing (McNeil and Sasse, 2016). Under RCP8.5, the Southern Ocean is exposed to the dual effects of undersaturation and hypercapnia (Hauck and Volker, 2015; Sasse et al., 2015; McNeil and Sasse, 2016). The *confidence* of these projections is *high* but for their timing it is *medium* due to model uncertainties.

The importance of the seasonal cycle is apparent when considering the year of onset of month-long and annual-mean undersaturation for the Southern Ocean under different scenarios for CMIP5 models: an abrupt change threshold is projected between RCP2.6 and RCP4.5/RCP8.5, with the latter two scenarios leading to the onset of pervasive mean annual undersaturation within 10 to 20 years of the onset of monthly undersaturation (Appendix 3.A.2.4, Table 3.A.2). By contrast, for RCP2.6 the area impacted by seasonal undersaturation is 0.2% of the RCP4.5/8.5 scenarios. The existence of this threshold is further supported by predictions based on the RCP8.5 scenario, that because of reduced buffering capacity, the onset of month-long hypercapnia in the Southern Ocean will occur around 2080, and that by 2100 almost the whole Southern Ocean will be impacted (McNeil and Sasse, 2016). This implies that under RCP4.5/8.5 scenarios, both calcification and organism physiology will be affected across Southern Ocean ecosystems (Sasse et al., 2015). Despite the importance of the seasonal cycle, recent studies highlight that interannual variability driven by large scale atmospheric modes (ENSO and SAM) should be included in the predictions for the onset of both undersaturation and hypercapnia (Conrad and Lovenduski, 2015). Although the *confidence* level for the onset of reduced buffering capacity and undersaturation is *high* to *very high*, the model projections for the timing of undersaturation and hypercapnia are still temporally and spatially uncertain so the overall *confidence* levels are *medium* to *high*.

3.2.3 Impacts on Marine Ecosystems

3.2.3.1 Arctic

The impacts of climate change on the polar ocean and cryosphere described in previous sections, presently have, and are projected to continue to have, significant implications for Arctic marine ecosystems, with consequences at different trophic levels both in the pelagic and benthic realm (Figure 3.4) (*high confidence*). Specifically, climate change is expected/projected to alter the distribution and properties of Arctic marine habitats with associated implications for the species composition, production and ecosystem structure and function (Moore et al., 2016; Frainer et al., 2017; Kaartvedt and Titelman, 2018) (*medium confidence*). These changes will modulate the Atlantic and Pacific Arctic gateways to the Arctic ecosystems (Mueter et al., 2017; Joli et al., 2018). The impacts of climate change on polar marine ecosystems are spatially heterogeneous between ecoregions (identified by Carmack et al. (2015); see map in Appendix 3.A, Figure 7) with respect to the rate and severity of change (*high confidence)*.

Climate change impacts on vertical fluxes and watermass layering may contribute to changes in benthic pelagic coupling (Kaartvedt and Titelman, 2018). In the few Arctic regions where data is sufficient to assess trends in biodiversity, the system level responses appear to be products of multiple interacting physical, chemical and biological processes (Frederiksen, 2017) (*medium confidence*). For instance, there is evidence that Chukchi Sea may be switching from dominating benthic to a more pelagic biomass production regime due to thinning sea ice and earlier ice retreat causing less new primary production to reach the sea floor and support benthos (Moore and Stabeno, 2015) (*low confidence*). Projected future reductions in summer sea ice, increased stratification in summer (Section 3.2.1.1), shifting currents and fronts and increased ocean temperatures (Section 3.2.1.2) and ocean acidification (Section 3.2.2.3) occur, they are expected to impact the future distribution of several marine fish and invertebrates (*high confidence*). Effects will be through direct and indirect pathways with severity of impacts being spatially heterogeneous and dependent on future emission scenarios.

Projected changes in seasonal ocean conditions based on the CMIP5 ensemble exhibit regional heterogeneity in the impacts of climate change (Appendix 3.A.2.6). Recent studies based on selected global models and those derived from CMIP5 indicate that the inflow systems of the northern Bering Sea, Chukchi Sea, Barents Sea and Kara Seas would be exposed to temperature increase (Renaud et al., 2015; Hermann et al., Submitted). It must be noted that there are significant limitations in CMIP5 projections of polar ocean temperature and sea ice on regional scales and that this lowers the confidence of the regional ecosystem impacts predictions.

3.2.3.1.1 Plankton and primary production

There is evidence that the combination of loss of sea ice, freshening, and regional stratification (Sections 3.2.1.1 and 3.2.1.2) has affected the timing, distribution and production of lower trophic level species (*high confidence*). Satellite data show that the decline in ice cover has resulted in a >30% increase in annual net primary production (NPP) in ice–free Arctic waters since 1998 (Arrigo and van Dijken, 2011; Bélanger et al., 2013; Arrigo and van Dijken, 2015; Kahru et al., 2016), a phenomenon corroborated by both in situ data (Stanley et al., 2015) and modelling studies (Vancoppenolle et al., 2013; Jin et al., 2016). Ice loss has also resulted in earlier phytoplankton blooms (Kahru et al., 2011) with blooms being dominated by larger-celled phytoplankton (Fujiwara et al., 2016). The longer open water season in the Arctic has also increased the incidence of autumn blooms, a phenomenon rarely observed in Arctic waters previously (Ardyna et al., 2017).

Thinner Arctic sea ice cover has led to the appearance of intense phytoplankton blooms that develop beneath first year sea ice (*medium confidence*). Observed in detail for the first time in the Arctic in 2011 (Arrigo et al., 2012), blooms of this size (1000s of km^2) and intensity (30 mg Chl-a m^{-3}) were previously thought to be restricted to the marginal ice zone and the open ocean where ample light reaches the surface ocean for rapid phytoplankton growth. Evidence shows that these blooms can thrive beneath sea ice in areas of reduced thickness, increased coverage of melt ponds (Arrigo et al., 2012; Arrigo et al., 2014; Zhang et al., 2015; Jin et al., 2016; Horvat et al., 2017), first year ridges at the snow-ice interface (Fernández-Méndez et al., 2018) and a large amount of cracks (lead fractions) in the ice (Assmy et al., 2017), although the latter has not changed significantly in the last three decades (Wang et al., 2016a).

The reduction in sea ice area and thickness in the Arctic Ocean appears to be indirectly impacting rates of NPP through increased exposure of the surface ocean to atmospheric forcing and these indirect impacts may increase in the future. Greater wind stress has been shown to increase upwelling of nutrients at the shelf break both over ice-free waters (Williams and Carmack, 2015) and a partial ice cover (Schulze and Pickart, 2012), leading to more new production (Williams and Carmack, 2015). At the same time, enhanced vertical stratification through the addition of freshwater at the ocean surface (Carmack et al., 2015) could decrease the upwelling of nutrients into surface waters (Capotondi et al., 2012; Nummelin et al., 2016), possibly reducing Arctic NPP in the future, especially in the central basin (Ardyna et al., 2017). It could also impact phytoplankton community composition and size structure, with small-celled phytoplankton becoming more dominant as nutrient concentrations in surface waters decline (Yun et al., 2015).

In addition to its impact on phytoplankton bloom dynamics, the decline in the proportion of multiyear sea ice and proliferation of a thinner first year sea ice cover may favor growth of microalgae within the ice due to increased light availability (*medium confidence*). Recent studies suggest that the contribution of sea ice algae to total Arctic NPP is higher now than values measured previously (Song et al., 2016), accounting for nearly 10% of total NPP (ice+water) and as much as 60% in places like the central Arctic (Fernández-Méndez et al., 2015).

Evidence suggests that these ongoing changes in NPP will impact the biogeochemistry and ecology of large parts of the Arctic Ocean (*high confidence*). In areas of enhanced nutrient availability and greater NPP, dominance by larger-celled microalgae increases vertical export efficiency from the surface downwards in both ice-covered (Boetius et al., 2013; Lalande et al., 2014; Mäkelä et al., 2017) and open ocean (Le Moigne et al., 2015) areas. However, because exported biomass production may be increasing in some areas but declining in others, the net impact may be small (Randelhoff and Guthrie, 2016) (see Appendix 3.A.2.5 for climate impacts on macroalgae). In addition, recent laboratory experiments suggest that Arctic phytoplankton assemblages may have the capacity to compensate for ocean acidification under a range of temperatures and P_{CO2} concentrations (Hoppe et al., 2018).

The phenology, magnitude and duration of zooplankton production and the zooplankton community composition in the Arctic are changing in response to increased water temperatures (Section 3.2.1) and the spatial pattern and timing of the ice algal and phytoplankton blooms (*medium confidence*). At the more southern boundaries of the Arctic such as the southeastern Bering Sea, warming conditions have led to a reduced production of large copepods and euphausiids, with consequences to recruitment of commercially important fish stocks such as walleye pollock (*Gadus chalcogrammus*) (Sigler et al., 2017; Kimmel et al., 2018). On more northern shelves, the increased open water period appears to have led to increases in large copepods over a 60–year period within the Chukchi Sea (Ershova et al., 2015) and in recent years also the Beaufort Sea (Smoot and Hopcroft, 2017), while in the Central Basins zooplankton biomass in general has increased (Hunt et al., 2014; Rutzen and Hopcroft, 2018).

Projections based on the SRES (IPCC, 2000) scenario A1B, suggest that large changes in the production, distribution and magnitude of the keystone copepods *Calanus finmarchicus* and especially *C. glacialis* in the Eurasian Arctic will occur towards the end of the century (Wassmann, 2015). Other studies have suggested that *C. glacialis* has, and should continue to, benefit from a warmer Arctic Ocean (Feng et al., 2018). Although, in the transition zone between Arctic and Atlantic water masses, *C. glacialis* may face increasing competition from *C. finmarchicus* (Dalpadado et al., 2016). In a study of Kongsfjorden, Spitsbergen (79°N) and adjacent waters, Dalpadado et al. (2016) concluded that that if projected warming trends persist, the Atlantic/boreal krill species will increase, while Arctic species, such as the amphipod *Thermisto libellula*, may decline (*low confidence)*.

Seasonal and spatial heterogeneity in the presence of undersaturated waters with respect to aragonite is expected in polar regions (Section 3.2.1.2.4) with marked differences in projected extent under different RCPs (Section 3.2.1.2.4). These changes will have associated impacts on calcifying zooplankton and pelagic mollusks (Luckman et al., 2014; Howes et al., 2015). The literature is mixed with respect to the projected severity of these impacts on pteropods, with recent Arctic studies demonstrating some resistance to the effects of acidification (but with unknown energetic costs) (Peck et al., 2016; Peck et al., 2018) (*low confidence*). Ocean acidification is expected to negatively impact survival of some crab and shellfish species in the future, however, current ocean conditions do not appear to have negatively impacted crab production in the Bering or Barents Seas (Mathis et al., 2015; Punt et al., 2015).

3.2.3.1.2 Benthic communities

There is evidence that earlier spring sea ice retreat and later autumn sea ice formation (Section 3.2.1.1) are changing the phenology of primary production with cascading effects on Arctic benthic community biodiversity and production (Link et al., 2013) (*medium confidence*). In the Barents Sea, evidence suggests that factors directly related to climate change (sea-ice dynamics, ocean mixing, bottom-water temperature change, ocean acidification, river/glacier freshwater discharge; sections 3.2.1.1 and 3.2.1.2) are impacting benthic species composition (Birchenough et al., 2015). Other human-influenced activities, such as commercial bottom trawling and introduction of non-indigenous species are also regarded as major drivers of observed and expected changes in benthic community structure (Johannesen et al., 2017), and may interact with climate impacts.

Over the last decade, a northward shift in the distribution of benthic species and subsequent changes in community composition have been detected in the northern Bering Sea (Grebmeier, 2012), Western Greenland (Renaud et al., 2015), and the Barents Sea (Kortsch et al., 2012; Kortsch et al., 2015) (*medium confidence*). Rapid and extensive structural changes in the rocky-bottom communities of two Arctic fjords in the Svalbard Archipelago have been documented during the period 1980 to 2010 and linked to gradually increasing seawater temperature and decreasing sea ice cover (Kortsch et al., 2012; Kortsch et al., 2015). Also, there is indication of declining benthic biomass in the northern Bering Sea (Grebmeier and Cooper, 2016) and southern Chukchi Sea (Grebmeier et al., 2015). It is unclear whether these rapid ecosystem changes will be tipping points for local ecosystems (Wassmann and Lenton, 2012). However, biomass of kelps have increased considerably in the intertidal to shallow subtidal in Arctic regions over the last 2 decades, connected to reduced physical impact by ice-scouring and increased light availability as a consequence of warming and concomitant fast-ice retreat (see also Appendix 3.A.3.5) (Kortsch et al., 2012; Bartsch et al., 2016; Paar et al., 2016) (*medium confidence*).

The production of Tanner and snow crab (*Chionoecetes bardi* and *C. opilio* respectively) and blue and red king crab (*Paralithodes platypus* and *P. camtschaticus* respectively) is also stressed by a complex suite of environmental drivers (Émond et al., 2015). In Newfoundland and Labrador waters and on the western Scotian Shelf, snow crab productivity has declined during a warm oceanographic regime (Mullowney et al., 2014; Zisserson and Cook, 2017). Contrary to this, snow crabs are expanding their distribution in the Barents Sea and commercial harvesting is rapidly increasing (Hansen, 2016; Lorentzen et al., 2018) (*high confidence*). Red king crab was intentionally introduced to the Barents Sea in the 1960s to support commercial fisheries in the Kola region and is now widely present in large numbers. Based upon thermal preferences, this species may potentially spread further north and east along the Euro-Arctic shelves within three decades or less (Christiansen et al., 2015) (*medium confidence*), exemplifying how thermal behaviour may drive the spreading of a marine invader under ocean warming (Box 3.3).

3.2.3.1.3 Fish

Changes in the spatial distribution and production of Arctic fish are best documented for ecologically and commercially important stocks in the Bering and Barents Seas (Box 3.3; Figure 3.4), while data is severely limited in most Arctic Ocean shelf regions. In the last decade, there is evidence that warm conditions have favoured fish production in the Barents Sea (*high confidence*), while being associated with reduced production of codfishes in the Bering Sea (*medium confidence*).

In the Barents Sea, heightened temperatures (Section 3.2.1.2) have expanded suitable feeding areas for boreal/subarctic species, which also contributed to increased Atlantic cod (*Gadus morhua*) production (Kjesbu et al., 2014). In contrast, Arctic species like polar cod (*Boreogadus saida*) are expected to be negatively affected by a shortened ice-covered season and reduced sea-ice extent through loss of spawning habitat and shelter, increased predatory pressure, reduced prey availability (Christiansen, 2017), and impaired growth and reproductive success (Nahrgang et al., 2014).Retrospective studies and laboratory experiments suggest that high lipid content zooplankton, important fish prey, may be less abundant in warm ocean conditions in the Bering Sea, resulting in reduced overwintering success of some Arctic and subarctic species (Heintz et al., 2013). Time series on responses of anadromous fish (including salmon) in the high Arctic is limited, although these stocks will also be exposed to a wide range of future stressors (Reist et al., 2016). There is some evidence that environmental variability influences the production of anadromous species such as Arctic char (*Salvelinus alpinus*), brown trout (*Salmo trutta*), and Atlantic salmon (*Salmo salar*) through its influence on environmental stressors governing growth and winter survival (Jensen et al., 2017).

Recent evidence supports previous findings that interannual and decadal environmental variability has impacted the productivity (growth and reproductive success) of some marine fish in the Barents and Bering Seas (*high confidence*). The annual production of fish stocks in high latitudes is governed by an array of complex processes that impact stocks differently throughout the first year of life; many of these processes are influenced by temperature variability (Ottersen et al., 2014; Szuwalski et al., 2014). Future climate change will affect and may disrupt such processes (*medium confidence*).

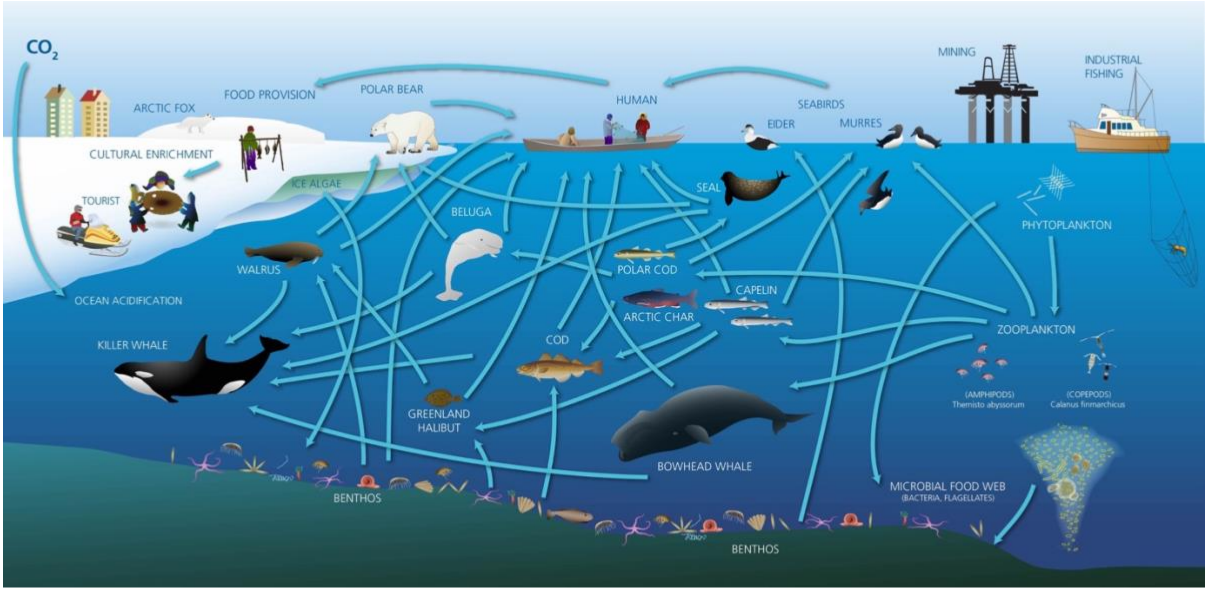


Figure 3.4: Schematic summary of the Arctic marine foodweb showing relationships important for ecosystem responses to climate change across different habitats (source CAFF (2017)).

The scope for adaptation of marine fish to a changing climate is uncertain, but knowledge is informed from previous biogeographic studies (Chernova, 2011; Lynghammar et al., 2013). The present niche partitioning between subarctic and Arctic pelagic fish species is expected to become more diffuse with potential negative impacts on cold adapted species such as Polar cod (Laurel et al., 2017; Logerwell et al., 2017) (*low confidence*). Winter ocean conditions in the high Arctic are projected to remain cold in most regions (Section 3.2.3.1), limiting the immigration of resident populations of subarctic species that spawn in positive temperatures onto the high Arctic shelves. Many demersal fish (groundfish) and invertebrates populations are constrained by the continental shelves and consequently they may not expand their habitat poleward beyond the shelf break. For instance, further expansion of Northeast Atlantic haddock (*Melanogrammus aeglefinus*) is expected to be limited to an eastward spreading along the Siberian shelf (Landa et al., 2014). Projected increases in summer temperature (Box 3.3) may open gateways to subarctic pelagic foragers in summer, particularly in the inflow regions of the Kara and Chukchi Seas, and the shelf regions of east and west Greenland. For example, the pelagic capelin (*Mallotus villosus*) are capable of entering the Polar Ocean, but they may be restricted in winter by availability of suitable spawning areas and lack of antifreeze proteins (Hop and Gjøsæter, 2013; Christiansen, 2017).

The indirect effects of changing ocean conditions include impacts on prey quality and distribution. Seasonal advection of pelagic prey may also allow feeding invasions to occur (Wassmann et al., 2015). Euphausiids and amphipods are a major food source for Arctic fishes, and changes in the composition of available prey towards smaller less energy-rich boreal species may have an impact on the feeding dynamics of these fish species (Dalpadado et al., 2016; Hunt et al., 2016). Further, large piscivorous and semipelagic boreal species may replace small-bodied benthivorous Arctic species as observed in the northern Barents Sea in the Atlantic sector, changing biogeography and ecosystem functioning (Box 3.3, Figure 1) (Fossheim et al., 2015; Frainer et al., 2017).

Only a few studies in the Barents and southeast Bering Seas have utilized current knowledge of mechanisms underlying fish responses to changing environmental conditions to project climate change impact on commercially important species. These studies show that under high emission scenarios, climate change will impact the future productivity of some commercially important marine fish stocks. Regional climate scenarios, derived from downscaled global climate scenarios, have been used to drive environmentally linked fish population models with temperature-specific growth and predation rates to project the impacts of climate change on the production of southeastern Bering Sea demersal fish (groundfish) (Hermann et al., 2016; Holsman et al., 2016; Ianelli et al., 2016; Hermann et al., Submitted). Holsman et al. (Submitted) contrasted future production of commercial fish stocks in the eastern Bering Sea under scenarios derived from projected downscaled high spatial and temporal resolution ocean habitats under RCP4.5 and 8.5. These scenarios projected future declines in in the abundance of walleye pollock, Pacific cod (*Gadus microcephalus*) and arrowtooth flounder (*Atheresthes stomias*) and fishery adaptation strategies were only effective in delaying the onset of decline. Based upon downscaled projections from GCMs and a spatially explicit Individual Based Model (IBM), Hedger et al. (2013) predicted increases in Atlantic salmon abundance, both in marine and freshwater stages in northern Norway (river Alta around 70°N).

3.2.3.1.4 Seabirds and marine mammals

Environmental alterations caused by global warming are resulting in phenological, behavioural, physiological, and distributional changes in Arctic marine mammal and seabird populations (Gilg et al., 2012; Post et al., 2013; Meier et al., 2014; Laidre et al., 2015; Barrett et al., 2017; Gall et al., 2017) (*high confidence*). These changes include responses to altered ecological interactions as well as direct responses to habitat degradation induced by especially loss of sea ice. Population responses to warming need not be linear, but may be particularly strong to abrupt warming events and associated regime shifts, as shown by black-legged kittiwakes (*Rissa tridactyla*) (Descamps et al., 2017).

Seabirds and marine mammals are mobile animals that respond to changes in the distribution of their preferred habitats and prey, by shifting their range, altering the timing or pathways for migration or prey shifting when this is feasible (Post et al., 2013; Lydersen et al., 2014; Gremillet et al., 2015; Kuletz et al., 2015; Laidre et al., 2015; Barrett et al., 2017). Changes in the location or availability of polar fronts, polynyas, tidal glacier fronts or ice edges have impacted where Arctic sea birds and marine mammals concentrate because of the influence these physical features have on productivity; traditionally these areas have been key foraging sites for top predators in the Arctic (Jay et al., 2012; deHart and Picco, 2015; Gremillet et al., 2015; Kuletz et al., 2015; Hunt et al., 2016; Hamilton et al., 2017; Hauser et al., 2017; Ramírez et al., 2017; Hunt et al., 2018).

In some species, shifts in distribution in response to changes in suitable habitat have been associated with increased mortality. Increased mortality rates of walrus (*Odobenus rosmarus)* calves have been observed during on-shore stampedes of unusually large herds, because Pacific walrus females are no longer able to haul out on ice over the shelf in summer due to the retraction of the southern ice edge into the deep Arctic Ocean (Kovacs et al., 2016). Shifts in the temporal and spatial distribution and availability of suitable areas of sea-ice for ice-breeding seals have occurred (Bajzak et al., 2011; Øigård et al., 2013) with increases in pup mortality and stranding in light ice years (Johnston et al., 2012; Soulen et al., 2013; Stenson and Hammill, 2014).

Climate impacts that reduce the availability of prey resources can negatively impact marine mammals (Asselin et al., 2011; Øigård et al., 2014; Hamilton et al., 2016a; Brown et al., 2017b; Choy et al., 2017). Evidence suggests that sea ice changes have increased the foraging effort required by ringed seals (*Pusa hispida*) in the marginal ice zone north of Svalbard (Hamilton et al., 2015) and resulted in diet shifts in coastal ringed seals in the same region (Hamilton et al., 2016b; Lowther et al., 2017). Ringed seals in Svalbard are using terrestrial haul-out sites during summer for the first time in observed history, following major declines in sea ice (Lydersen et al., 2017); an example of an adaptive behavioural response to extreme habitat changes. Sea ice related changes in the export of production to the benthos (Section 3.3.3.1) and associated changes in the benthic community (Section 3.4.1.1.2) may impact marine mammals dependent on benthic prey (e.g., walruses and gray whales, *Eschrichitus robustus*) (Brower et al., 2017; Udevitz et al., 2017; Szpak et al., 2018).

Changes in the timing, distribution and thickness of sea ice and snow (Sections 3.2.1.1, 3.4.1.1) have been linked to phenological shifts, and changes in distribution, denning, foraging behaviour and survival rates of polar bears (*Ursus maritimus*) (Derocher et al., 2011; Hamilton et al., 2017; Olson et al., 2017; Escajeda et al., 2018). Less ice is also driving polar bears to travel over greater distances and swim more than previously both in offshore and in coastal areas, which can be dangerous for young cubs (Aars et al., 2017; Durner et al., 2017; Pilfold et al., 2017; Lone et al., 2018; Rode et al., 2018). Cumulatively, changes in sea ice patterns are driving demographic changes in polar bears, including declines in some populations where sea ice reductions are notable (Lunn et al., 2016; McCall et al., 2016). However, some polar bear populations are stable or increasing (Voorhees et al., 2014), even with regional declines in sea ice. This is because protective management measures have been successful in allowing severely depleted populations to recover despite habitat degradation or because new food sources, such as carrion from killer whale (*Orcinus orca*) takes of bowhead whales (*Balaena mysticetus*) are becoming available to polar bears in some regions (Galicia et al., 2016; Stapleton et al., 2016). Changes in the spatial distribution of polar bears and killer whales can have top-down effects on other marine mammal prey populations (Reinhart et al., 2013; Øigård et al., 2014; Breed et al., 2017; Smith et al., 2017a).

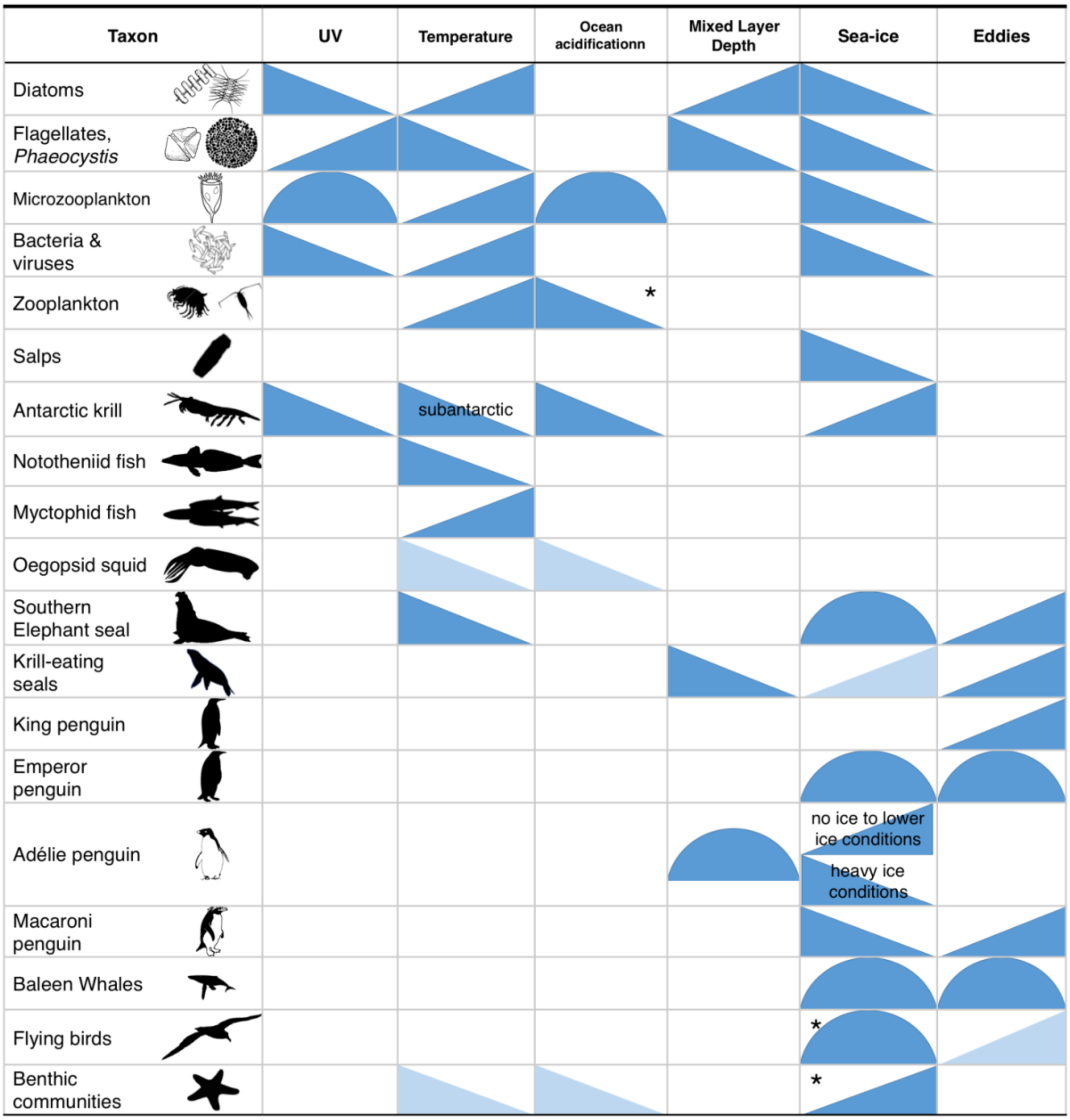
In recent decades, several studies from different parts of the Arctic show evidence that changes in seabird diets (Dorresteijn et al., 2012; Divoky et al., 2015; Kokubun et al., 2018; Vihtakari et al., 2018), reproductive success and body condition (Gaston et al., 2012; Provencher et al., 2012; Gaston and Elliott, 2014), and local seabird species composition (Gall et al., 2017) are occurring in response to changes in sea surface temperature and sea ice dynamics and their impact on the distribution and abundance of seabird prey (*medium confidence*).

Several studies from different parts of the Arctic show evidence that changing temperatures impact seabirds diets (Dorresteijn et al., 2012; Divoky et al., 2015; Kokubun et al., 2018; Vihtakari et al., 2018), reproductive success and body condition (Gaston et al., 2012; Provencher et al., 2012; Gaston and Elliott, 2014). Recent studies also show changes in sea surface temperature and sea ice dynamics impact on the distribution and abundance of seabird prey with cascading impacts on local seabird species composition (Gall et al., 2017), nutritional stress, and decreased reproductive output (Dorresteijn et al., 2012; Kokubun et al., 2018) and survival (Renner et al., 2016; Hunt et al., 2018). In the western Beaufort Sea, increasing sea surface temperature and loss of sea ice negatively affected the black guillemot (*Cepphus grylle mandtii*), an ice-obligate diving seabird, by changing its access to juvenile Arctic cod (Divoky et al., 2015).

3.2.3.2 Southern Ocean

Marine ecosystem dynamics in the Antarctic region are dominated by the ACC and its frontal systems (Cross-Chapter Box 5), subpolar gyres, polar seasonality, the annual advance and retreat of sea ice (Section 3.2.1.1), and the supply of limiting micronutrients for productivity (most commonly iron). Antarctic krill (*Euphausia superba*) play a central role in Southern Ocean foodwebs as grazers and as prey items for fish, squid, marine mammals and seabirds (Schmidt and Atkinson, 2016; Trathan and Hill, 2016). This is in part due to their abundance and circumpolar distribution, although the abundance and importance of this species varies between different regions of the Southern Ocean (Larsen et al., 2014; Siegel, 2016; McCormack et al., In review). Recent work has characterised the nature of habitat change for Southern Ocean biota at regional and circumpolar scales (Constable et al., 2014; Gutt et al., 2015; Hunt et al., 2016; Murphy et al., 2016; Gutt et al., 2017; Trebilco et al., In review), and the direct responses of biota to these changes (Constable et al., 2014) (Table 3.1). These findings indicate that overlapping changes in key ocean and sea-ice habitat characteristics (temperature, sea-ice cover, ice-berg scour, mixed layer depth, aragonite under-saturation; sections 3.2.1 and 3.2.2) will be important in determining future states of Southern Ocean ecosystems (Constable et al., 2014; Gutt et al., 2015) (*medium confidence*). Indirect responses to physical change remain less well characterized because they are numerous and because it is challenging to determine the relative strength of positive and negative feedbacks which dictate the direction of such effects. Important advances have also been made in (i) identifying key variables to detect and attribute change in Southern Ocean ecosystems, as part of long-term circumpolar modelling designs (Constable et al., 2016), and (ii) refining methods for using sea ice predictions from global climate models in ecological studies and in ecosystem models for the Southern Ocean (Cavanagh et al., 2017).

Table 3.1: Summary of known direct responses of biota to changes in physical parameters in Antarctica and the Southern Ocean (based on Constable et al. (2014); Constable et al. (2017)). UV = ultraviolet radiation. Acidification includes altered carbonate chemistry and pH. Sea-ice includes consideration of thickness, concentration, and extent without differentiating the factor/s causing change in each group of organisms. An upwards sloping triangle indicates a positive relationship (increase in the physical variable is expected to cause an increase in the taxon). A downwards sloping triangle indicates a negative relationship (increase in the physical variable is expected to cause a decline in the taxon). A humped symbol indicates a non-linear response (that can be positive or negative). Lighter shaded symbols represent uncertain responses (where the evidence is equivocal). These symbols do not indicate population responses to physical change, but show individual responses to physical drivers. As physical factors vary in their direction of change between different regions of the Southern Ocean, the responses in this table are used to interpret what specific directions of change may mean for the biota in a region. Indirect responses to physical parameters are addressed in the main text and are too numerous to capture in this table.



Notes:
* Indicates variations to the original version of the table, derived from (Jenouvrier et al., 2005; Olivier et al., 2005; Barnes and Souster, 2011; Bednaršek et al., 2012; Gardner et al., 2018b)

3.2.3.2.1 Phytoplankton and primary production

Changes in column-integrated phytoplankton biomass for the Southern Ocean are coupled with changes in the spatial extent of ice-free waters, suggesting little overall change in biomass per area at the circumpolar scale (Behrenfeld et al., 2016) (*low confidence*). Arrigo et al. (2008) also report no overall trend in remotely sensed column-integrated primary production south of 50°S for the period between 1998 and 2006, inclusive (*low confidence*). At a regional scale, local-scale forcings (e.g., retreating glaciers and topographically steered circulation) and stratification are key determinants of phytoplankton bloom dynamics at coastal stations on the Western Antarctic Peninsula (Kim et al., 2018) (*medium confidence*). Schofield et al. (2017) report a five-fold range of interannual variability in water column-integrated chlorophyll stocks, overlaid with a significant increase in the seasonal mixed-layer chlorophyll inventory over twenty years of local-scale observations from the West Antarctic Peninsula. The phenology of Southern Ocean phytoplankton blooms in this region may also be shifting to earlier in the growth season (Arrigo et al., 2017b) (*low confidence*). However, the effect of climate change on Southern Ocean primary production is difficult to determine given that the length of time series data is insufficient (less than 30 years) to enable the climate change signature to be detected and attributed; and that, even when records are of sufficient length, data trends are often reported as being driven by climate change when they are due to a combination of climate change and variability (*high confidence*).

Experimental studies on the effect of environmental drivers on phytoplankton growth rate indicate an important role for temperature and iron supply in polar waters (Xu et al., 2014; Hutchins and Boyd, 2016) (*low confidence*). Recent studies on coastal phytoplankton indicate a detrimental effect of acidification on phytoplankton communities (Hancock et al., 2017; Deppeler et al., 2018; Westwood et al., 2018) (*medium confidence*). McMinn (2017) reviewed the effects of acidification on sea-ice algae, during laboratory manipulations lasting days to weeks, and reported that in general acidification caused no detrimental effects to the study organisms. In situ experiments also revealed a tolerance to acidification, and as for the laboratory studies provided evidence of either no change in metabolic rates or increased rates (*medium confidence*). Model projections of trends in primary production in the Southern Ocean due to climate change from (Leung et al., 2015) are summarized in Table 3.2.

Table 3.2: Model projections of trends due to climate-change driven alteration of phytoplankton properties under RCP8.5 from 2006–2100 across three zones of the Southern Ocean, modified from Leung et al. (2015). Acidification was not reported as an important driver in this modelling experiment.

Zonal Band	Predicted change in phytoplankton biomass	Drivers	Mechanisms
40°S–50°S	↑	Higher underwater irradiance; more iron supply	Shallowing of the summertime mixed layer depth (which alleviates light limitation); change in iron supply mechanism
50°S–65°S	↓	Lower underwater irradiance	Combination of deeper summertime mixed layer depth along with decreased summertime incident radiation due to increased total cloud fraction
S of 65°S	↑	More iron supply and higher underwater irradiance; temperature	Melting of sea–ice Warming ocean

3.2.3.2.2 Krill and zooplankton

Previously reported declines in Antarctic krill abundance in the South Atlantic sector (Atkinson et al., 2004; Larsen et al., 2014) may reflect a sudden, discontinuous change following an episodic period of anomalous peak abundance for this species (Loeb and Santora, 2015) rather than an ongoing decline (*medium confidence*). Recent analyses have not detected trends in long-term krill abundance in the South Atlantic sector (Fielding et al., 2014; Kinzey et al., 2015; Steinberg et al., 2015; Cox et al., 2018). Nevertheless, given its dependence on sea ice habitats, the Antarctic krill population may already have changed (*medium confidence*) and will be subject to further alterations (*high confidence*).

The distribution of Antarctic krill is expected to change under future climate change because of changes in the location of the optimum conditions for krill growth and recruitment (Melbourne-Thomas et al., 2016; Piñones and Fedorov, 2016; Meyer et al., 2017; Klein et al., 2018; Trebilco et al., In review). Based on empirical evidence for the relationship between temperature and krill growth and recruitment, the optimum conditions for krill are predicted to move southwards, with the decreases most apparent in the areas with the most rapid warming (Hill et al., 2013; Piñones and Fedorov, 2016) (Section 3.2.1.2.1) (*medium confidence*). The predicted impacts of temperature changes and ocean acidification on Antarctic krill are not homogeneously distributed; the greatest reductions in krill are predicted for the southwest Atlantic/Weddell Sea region (Kawaguchi et al., 2013; Piñones and Fedorov, 2016) (*low confidence*), which is the area of highest current krill concentrations, contains important foraging grounds for krill predators, and is also the area of operation of the krill fishery. Modelled effects of warming on krill growth in the Scotia Sea and northern Antarctic Peninsula region resulted in reductions in total krill biomass under both RCP2.6 and 8.5, with a 25% chance of krill biomass falling below 75% of a reference scenario (no fishing or climate change) under RCP8.5 (Klein et al., 2018) (*low confidence*). Projections from a food web model for the West Antarctic Peninsula under simple scenarios for change in open water and sea ice associated primary production from 2010 to 2050 indicate a decline in krill biomass with contemporaneous increases in the biomass of gelatinous salps (Suprenand and Ainsworth, 2017).

Current understanding of climate change effects on Southern Ocean zooplankton is largely based on observations and predictions from the South Atlantic and the West Antarctic Peninsula. Comparison of the mesozooplankton community in the southwestern Atlantic sector between 1926 and 1938 and 1996–2013 showed no evidence of change despite surface ocean warming (Tarling et al., 2017) (*medium confidence*). These results suggest that predictions of distributional shifts based on temperature niches may not reflect the actual levels of thermal resilience of key taxa. Sub-decadal cycles of macrozooplankton community composition adjacent to the West Antarctic Peninsula are strongly linked to climate indices, with evidence of increasing abundance for some species over the period from 1993 to 2013 (Steinberg et al., 2015) (*low confidence*). Pteropods are vulnerable to effects of acidification, and new evidence indicates that eggs released at high pCO_2 lack resilience to ocean acidification in the Scotia Sea region (Manno et al., 2016) (*medium confidence*).

3.2.3.2.3 Fish

Many Antarctic fish have a narrow thermal tolerance as a result of physiological adaptations to cold water (antifreeze glycopeptides that prevent body fluids from freezing and decreased haematocrit and haemoglobin concentrations; (Beers and Sidell, 2011; Mintenbeck, 2017), which makes them vulnerable to the effects of increasing temperatures (Mueller et al., 2012). Increasing water temperatures may displace icefish (family Channichthyidae) in marginal habitats as they lack haemoglobin and are unable to adjust blood parameters to an increasing oxygen demand (Mintenbeck et al., 2012) (*low confidence*). The Antarctic silverfish (*Pleuragramma antarcticum*) is an important prey species in some regions of the Southern Ocean, and has an ice-dependent life cycle (Mintenbeck et al., 2012; Vacchi et al., 2012). Documented declines in the abundance of this species in some parts of the West Antarctic Peninsula may have consequences for associated food webs (Parker et al., 2015; Mintenbeck and Torres, 2017) (*low confidence*).

Myctophids and toothfish are important fish groups from both a food web (myctophids) and fishery (toothfish) perspective. Ocean warming (Section 3.2.1.3.4) is expected to cause southward shifts in the distributions of myctophid fish species and could also result in isolated populations restricted to island shelves becoming locally extinct, if they are unable to adapt to warmer ocean temperatures (Constable et al., 2014) (*low confidence*). There is no evidence for effects of climate change on the two species of toothfish that are found in the Southern Ocean; Patagonian and Antarctic toothfish (*Dissostichus eleginoides* and *D. mawsoni*). There is *limited evidence* that recruitment is inversely correlated with sea surface temperature for Patagonian toothfish at South Georgia (Belchier and Collins, 2008) (*medium confidence*). Given differences in temperature tolerances for Patagonian toothfish (with a wide temperature tolerance) and Antarctic toothfish (limited by a low tolerance for water temperatures above 2°C), the latter may be faced with reduced habitat and potential competition with southward-moving Patagonian toothfish under climate change (Mintenbeck, 2017) (*very low confidence*).

3.2.3.2.4 Seabirds and marine mammals

Population trends for Antarctic seabirds and marine mammals vary within and among Southern Ocean sectors (as defined by Bost et al. (2009); Gutt et al. (2015); Hunt et al. (2016); Murphy et al. (2016); Gutt et al. (2017); Trebilco et al. (In review)) and reflect the different drivers affecting them, particularly sea-ice extent and food availability (*high confidence)* across regions (Section 3.2.1.1.1). The predictability of foraging grounds and ice-coverage are associated with variations in climate (Crocker et al., 2006; Baez et al., 2011; Dugger et al., 2014; Abrahms et al., 2017; Youngflesh et al., 2017) (Section 3.2.1.1) and are the main drivers of observed population changes of Southern Ocean top predators (*high confidence*) (Descamps et al., 2015; Jenouvrier et al., 2015; Sydeman et al., 2015; Abadi et al., 2017; Bjorndal et al., 2017; Fluhr et al., 2017; Hinke et al., 2017a; Hinke et al., 2017b; Pardo et al., 2017). The suitability of breeding habitats and the location of environmental features that facilitate the aggregation of prey are also influenced by climate change and in turn influence the distribution in space and time of marine mammals and birds (Bost et al., 2015; Kavanaugh et al., 2015; Hindell et al., 2016; Santora et al., 2017) (*medium confidence*). Finally, biological parameters (reproductive success, mortality, fecundity, condition), life history traits, morphological, physiological and behavioural characteristics of top predators in the Southern Ocean, as well as their patterns of activity (migration, distribution, foraging, reproduction) are also changing as a result of climate change (Whitehead et al., 2015; Braithwaite et al., 2015a; Seyboth et al., 2016; Hinke et al., 2017b) (*high confidence*).

Population changes associated with climate change for Antarctic penguins include both increases (for Gentoo penguins, *Pygoscelis papua*) (Lynch et al., 2013; Dunn et al., 2016; Hinke et al., 2017a), and decreases (for Adélie (*P. adeliae)* Chinstrap (*P. antarctica)* King (*Aptenodytes patagonicus*) and Emperor (*A. forsteri*) penguins) (Trivelpiece et al., 2011; LaRue et al., 2013; Jenouvrier et al., 2014; Bost et al., 2015; Southwell et al., 2015; Younger et al., 2015; Cimino et al., 2016) (*high confidence*). Youngflesh et al. (2017) suggest that population shifts observed in Adélie penguins are a result of strong interannual environmental variability in good and bad years for prey and breeding habitat rather than climate-change (*low evidence*). New evidence suggests that present Emperor penguin population estimates should be evaluated with caution based on the existence of breeding colonies yet to be discovered/confirmed (Ancel et al., 2017) as well as studies that draw conclusions based on trend estimates from single colonies (Kooyman and Ponganis, 2017). Evidence for climate change impacts on Antarctic flying birds indicates that contraction of sea ice, increases in sea surface temperatures and extreme events (snow storms) can reduce breeding success and population growth rates in some species (Southern Fulmars (*Fulmarus glacialoides*), Antarctic Petrels (*Thalassoica Antarctica*) and Black-browed albatrosses (*Thalassarche melanophris*) (Jenouvrier et al., 2015) (Descamps et al., 2015; Pardo et al., 2017) (*low confidence)*.

Local and regional-scale oceanographic features (Section 3.2.1.2) and bathymetry controlling prey aggregations affect the ecological responses and biological traits of marine mammals in the Southern Ocean (Lyver et al., 2014; Bost et al., 2015; Jenouvrier et al., 2015; Whitehead et al., 2015; Cimino et al., 2016; Seyboth et al., 2016; Hinke et al., 2017a; Pardo et al., 2017) (*high confidence*) and *likely* explain most of observed population shifts (Kavanaugh et al., 2015; Hindell et al., 2016; Gurarie et al., 2017; Santora et al., 2017). Decadal climate cycles affect access to mesopelagic prey by Southern elephant seals (*Mirounga leonina*) in the Indian Sector of the Southern Ocean and breeding females are excluded from highly productive continental shelf waters in years of increased sea ice extent and duration (Hindell et al., 2016) (*medium confidence)*. To date there is no unified global estimate of the abundance of Antarctic pack ice seal species (Ross seals (*Omatophoca rossi)*, Crabeater seals (*Lobodon carciniophaga)*, Leopard seals (*Hydrurga leptonyx)* and Weddell seals (*Leptonichotes weddellii)*) as a reference point for understanding climate change impacts on these species (Constable et al., 2017). Analysis of long-term data suggests a genetic component to adaptation to climate change (*low confidence*) in Antarctic fur seals (*Arctocephalus gazella*, Forcada and Hoffman (2014) and pigmy blue whales (*Balaenoptera musculus brevicauda*, Attard et al. (2015)).

Population trends of migratory baleen whales have been associated with krill abundance in the Atlantic and Pacific sectors of the Southern Ocean which is reflected in increased reproductive success, body condition and energy allocation (milk availability and transfer) to calves (Braithwaite et al., 2015a; Braithwaite et al., 2015b; Seyboth et al., 2016) (*high confidence*). Whale biological parameters reflect the connectivity of environmental conditions between whale foraging (Southern Ocean) and breeding grounds (lower latitudes).

3.2.3.2.5 Pelagic and benthic ecosystems

This section assesses the impacts of ocean and sea ice changes on pelagic and benthic ecosystem structure, dynamics and biodiversity (see also Figure 3.5). The ecological impacts of loss of ice shelves and retreat of coastal glaciers around Antarctica are assessed in Section 3.3.3. Recent syntheses of Southern Ocean ecosystem structure and function recognise the importance of at least two dominant energy pathways in pelagic foodwebs – a short trophic pathway transferring primary production to top predators via krill, and at least one other pathway that moves energy from smaller phytoplankton to top predators via copepods and small mesopelagic fishes – and indicate that the relative importance of these pathways will change under climate change (Murphy et al., 2013; Murphy et al., 2016; Constable et al., 2017; McCormack et al., In review) (*medium confidence*). Using an ecosystem model, Klein et al. (2018) found that the effects of warming on krill growth off the Antarctic Peninsula and in the Scotia Sea translated to increased risks of krill predator populations, particularly penguins, declining to less than 75% of modelled abundances in reference scenarios (without warming or fishing) under both RCP2.6 and RCP8.5. The relative importance of different energy pathways in Southern Ocean foodwebs has important implications for resource management, in particular the management of krill and toothfish fisheries by the Commission for the Conservation of Antarctic Marine Living Resources (CCAMLR) (Murphy et al., 2016; Constable et al., 2017) (see Sections 3.2.4.1.2, 3.5.4.2.2).

Bentho-pelagic coupling and vertical energy flux will influence marine ecosystem responses to climate change. New modelling approaches have recently become available to better capture these relationships at large spatial scales (Jansen et al., 2017). Griffiths et al. (2017a) use species distribution modelling for 963 benthic invertebrate species in the Southern Ocean to consider distribution changes under RCP8.5 for 2099. Their results suggest that 79% of Antarctica’s endemic species will face a reduction in suitable temperature habitat (an average 12% reduction) over the current century (*low confidence*). Predicted reductions in the number of species are most pronounced for the West Antarctic Peninsula and the Scotia Sea region (Griffiths et al., 2017a).

Carbon uptake and storage by Antarctic benthic communities is predicted to increase with sea ice losses, because across-shelf growth gains from longer algal blooms outweigh ice scour mortality in the shallows (Barnes, 2017) (*low confidence*). Communities in shallow water habitats mostly consist of dark-adapted invertebrates, and rely on sea ice to create low-light marine environments. Increases in the amount of light reaching shallow seabed under climate change may result in ecological regime shifts, in which invertebrate-dominated communities are replaced by macroalgal beds (Clark et al., 2015; Clark et al., 2017) (*low confidence*).

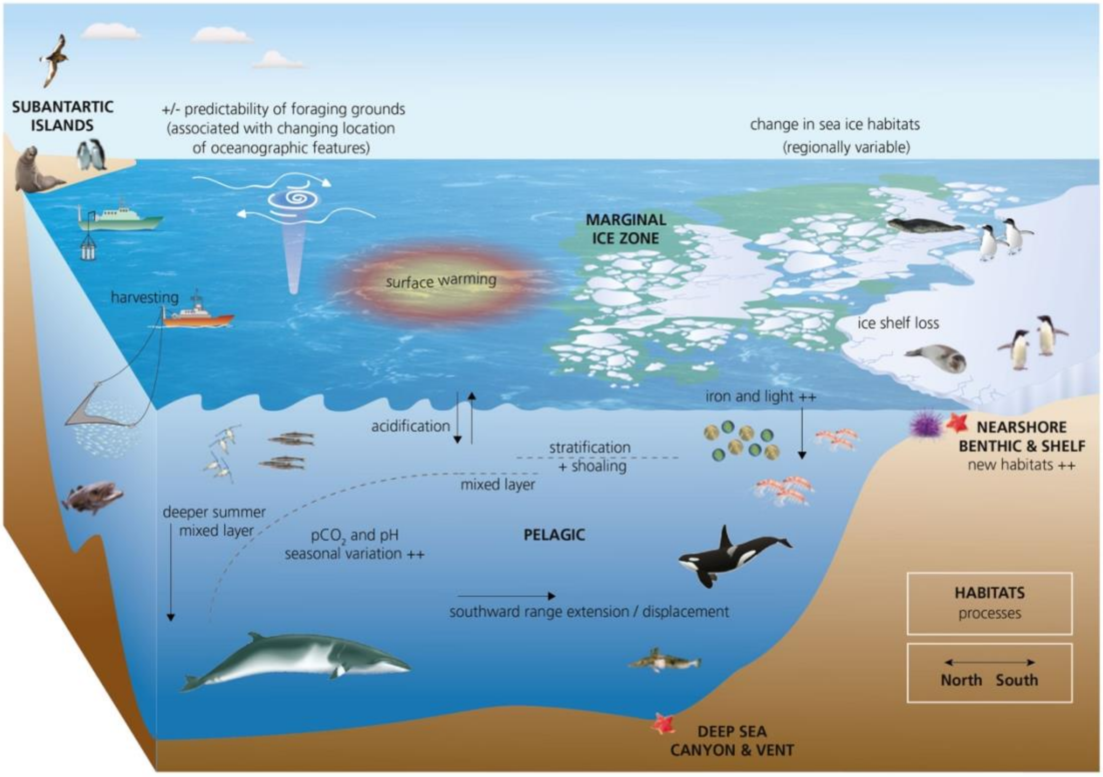


Figure 3.5: Schematic summary of key processes determining ecosystem responses to climate change across different habitats in the Southern Ocean.

3.2.4 Impacts on Social-Ecological Systems

3.2.4.1 Fisheries

3.2.4.1.1 Arctic

Arctic fisheries are economically and socially important. Large commercial fisheries exist off the coasts of Greenland and in the Barents and Bering Seas (Holsman et al., 2018; Peck and Pinnegar, 2018). The target species for these commercial fisheries include gadids, flatfish, herring, red fish (*Sebastes* sp.), salmonids, and capelin, among others. Fisheries in other Arctic regions are relatively small scale, locally operated and they target a limited number of species (Reist, 2018). Although these fisheries are small, they are also of considerable cultural, economic, and subsistence importance to local communities (Section 3.5.4.1).

Projecting the impacts of climate change on marine fisheries is inextricably intertwined with response scenarios regarding risk tolerance in future management of marine resources, advancements in fish capture technology, and markets drivers (e.g., local and global demand, emerging product lines, competition, processing efficiencies and energy costs) (Groeneveld et al., 2018). Seasonal and interannual variability in ocean conditions influences product quality, and costs of fish capture (Haynie and Pfeiffer, 2012) (see also Table 3.7). Past experience suggests that barriers to diversification may limit the portfolio of viable target fisheries available to both small and large scale fisheries (Ward et al., 2017). If managed sustainably, some Arctic fisheries may be able to adapt to moderate future warming (European Parliament's Committee on Fisheries, 2015). Past performance also suggests that high latitude fisheries have been resilient to changing environmental and market drivers. For example, the Norwegian cod fishery has exported dried cod over an unbroken period of more than thousand years (Barrett et al., 2011), reflecting the resilience of the northern Norwegian cod fisheries to historic climate variability (Eide, 2017).

Climate change will affect the spatial distribution and productivity of some commercially important marine fish and shellfish under most RCPs (Section 3.2.3.1) with associated impacts on the distribution and economic viability of commercial fisheries (*high confidence*). Model projections indicate that expansions in suitable habitat for subarctic species and increased production of planktonic prey due to increasing temperatures and ice retreat, will continue to support commercially important fisheries in the Atlantic regions (Lam et al., 2016; Eide, 2017; Haug et al., 2017; Peck and Pinnegar, 2018) (Section 3.2.3.1.3 and Box 3.3). However, recent studies in the Bering Sea suggest that future fish production will also depend on how climate change and ocean acidification will alter: the quality, quantity and availability of suitable prey; the thermal stress and metabolic demands if resident fish; and species interactions (Section 3.2.3.1.3) suggesting that the future of commercial fisheries in Arctic regions is uncertain (Holsman et al., 2018). It is also uncertain whether future autumn and winter ocean conditions will be conducive to the establishment of resident overwintering spawning populations that are large enough to support sustainable commercial fishing operations at higher latitude shelf regions of the Arctic (Section 3.2.3.1).

3.2.4.1.2 Southern Ocean

This section examines climate change impacts on Southern Ocean fisheries for Antarctic krill and finfish. Management of these fisheries by the CCAMLR and responses to climate change are discussed in Section 3.5.2.1. The main Antarctic fisheries are for Antarctic krill, and for Antarctic and Patagonian toothfish; in 2016 the reported catches for these species were approximately 260 thousand tons for krill (CCAMLR, 2017b) and 11 thousand tons for Antarctic and Patagonian toothfish combined (CCAMLR, 2017a). The fishery for Antarctic krill in the southern Atlantic sector and the northern West Antarctic Peninsula (together the current area of focus for the fishery) has become increasingly concentrated in space over recent decades, which has raised concern regarding localised impacts on krill predator (Hinke et al., 2017a). The krill fishery has also changed its peak season of operation. In the early years of the fishery, most krill were taken in summer and autumn, with lowest catches being taken in spring. In recent years the lowest catches have occurred over summer, catches have peaked in late autumn, and very little fishing activity has occurred in spring (Nicol and Foster, 2016). Some of these temporal and spatial shifts in the fishery over time have been attributed to reductions in winter sea-ice extent in the region (Kawaguchi et al., 2009) (*medium confidence*). Recent increases in the use of krill catch to produce krill oil (as a human health supplement) has also led to vessels concentrating on fishing in autumn and winter when krill are richest in lipids (Nicol and Foster, 2016) (*medium confidence*). Available evidence regarding future changes to Antarctic krill populations (Section 3.2.3.2.2) indicates that the impacts of climate change will be most pronounced in the areas that are currently most important for the Antarctic krill fishery; the Scotia Sea and the northern tip of the Antarctic Peninsula. Major future changes in the krill fishery itself are expected to be driven by global issues external to the Southern Ocean, including conservation decision making, socio-economic drivers and geopolitics.

There is *limited evidence* available regarding the consequences of climate change for Southern Ocean finfish fisheries. Lack of recovery of mackerel icefish (*Champsocephalus gunnari*) after cessation of fishing in 1995 has been related to anomalous water temperatures (~2°C increase related to a strong El Niño) in the subantarctic Indian Ocean and to availability of krill prey in the Atlantic region (Mintenbeck, 2017) (*low* *confidence*). Differences in temperature tolerance of Patagonian and Antarctic toothfish described in Section 3.2.3.2.3 may have implications for future fisheries of these two species.

3.2.4.2 Tourism

Reductions in sea ice have facilitated an increase in marine and cruise tourism opportunities across the Arctic related directly to an increase in accessibility (Dawson et al., 2014; Dawson et al., 2017) (*very high confidence*). While not strictly 'polar', Alaska attracts the highest number of cruise passengers annually at just over one million; Svalbard attracts 40,000–50,000; Greenland 20,000–30,000; and Arctic Canada 3,500–5,000 (Dawson et al., 2017). Compared to a decade ago, there are more cruises on offer, ships travel further in a single season, larger vessels with more passenger berths are in operation, purpose-built polar cruise vessels are being constructed, and private pleasure craft are appearing in greater frequency (Lasserre and Têtu, 2015; Johnston et al., 2017; Dawson et al., 2018). In Antarctica, almost 37,000 predominantly ship-borne tourists visited in 2016/17. Due to accessibility and convenience, these tourism operations are mostly based around the few ice-free areas of Antarctica, concentrated on the Antarctic Peninsula (Pertierra et al., 2017).

There is *high confidence* that Arctic cruise tourism will continue to grow over the coming decade (Johnston et al., 2017) due to continued sea ice reduction and 'anticipated' climate change-related perceptions of increased navigability. For example, Canada's Northwest Passage (southern route), which only saw occasional cruise ship transits in the early 2000s is now reliably accessible during the summer cruising season and as a result has experienced a doubling and quadrupling of cruise and pleasure craft activity over the past decade (Johnston et al., 2017; Dawson et al., 2018). The anticipated implications of future climate change have also led to the emergence of a niche tourism market known as 'last chance tourism' – whereby tourists explicitly seek to experience vanishing landscapes or seascapes, and natural and social heritage in the Arctic and Antarctic before they disappear (Lemelin et al., 2010; Lamers et al., 2013).

Increases in polar cruise tourism pose risks and opportunities related to development, education, safety (including search and rescue), security, and environmental sustainability (Johnston et al., 2012a; Johnston et al., 2012b; Stewart et al., 2013; Dawson et al., 2014; Lasserre and Têtu, 2015; Stewart et al., 2015). In the Arctic, there are also risks and opportunities related to employment, health and well-being, and the commodification of culture (Stewart et al., 2013; Stewart et al., 2015). The biodiversity supported by ice-free areas, particularly those on the Antarctic Peninsula, has been identified as being particularly vulnerable to the introduction of terrestrial alien species (Hughes et al., 2015; Duffy et al., 2017; Lee et al., 2017a) (see Box 3.3) as well as to the direct impacts of the humans (e.g., through trampling Pertierra et al., 2017). Because the sector relies on a set of regulations that apply to all types of maritime shipping, yet cruise ships purposefully travel off regular shipping corridors, a need for appropriate governance regimes, specialized infrastructure, and focused policy attention has been identified (Dawson et al., 2014; Pashkevich et al., 2015; Dawson et al., 2016; Dawson et al., 2017). Private pleasure craft remain almost completely unregulated, and will pose unique risks in the future (Johnston et al., 2017).

3.2.4.3 Transportation

The Arctic is reliant on marine transportation for the import of food, fuel, and other goods, while the global appetite for maritime trade and commerce through the Arctic (including community re-supply, mining and resource development, tourism, fisheries, cargo, research, and military and icebreaking, etc.) is increasing as the region becomes more accessible because of reduced sea ice cover. There are four potential Arctic international trade routes: the Northwest Passage, the Northern Sea Route, Arctic Bridge, and Transpolar Sea Route. All of these routes offer significant trade benefits because they provide substantial distance savings compared to traditional routes via the Suez or Panama Canals.

There is *very high confidence* that climate change driven reductions in Arctic sea ice is a main driver of increased Arctic shipping activity over the past decade (Pizzolato et al., 2014; Eguíluz et al., 2016; Pizzolato et al., 2016; Dawson et al., 2018), with further influence from non-environmental factors such as natural resource development, regional trade, geopolitics, commodity prices, global economic and social trends, national priorities, tourism demand, ship building technologies, and insurance costs (Lasserre and Pelletier, 2011; Têtu et al., 2015; Dawson et al., 2017). It is projected that shipping activity will continue to rise across the Arctic as northern routes become increasingly accessible, although the influence of potential changes to insurance premiums are not clear (Stephenson et al., 2011; Smith and Stephenson, 2013; Barnhart et al., 2015; Melia et al., 2016). The Northern Sea Route is expected to be more viable than other routes, considering investments in infrastructure and favourable sea ice dynamics. In comparison, the Northwest Passage and Arctic Bridge have limited port and marine transportation infrastructure, limited soundings and incomplete hydrographic charting, and challenging sea ice conditions (Stephenson et al., 2013; Andrews et al., 2018). These conditions, together with limited search and rescue capacity and remote and harsh geography compound risks from shipping activity across the entire region (Dawson et al., 2017).

Projected changes to Arctic shipping activities will have significant socio-economic and political implications related to safety (marine accidents, local accidents, ice as a hazard), security (trafficking, terrorism), and environmental and cultural sustainability (invasive species, release of biocides, chemicals and other waste, marine mammal strikes, fuel spills, air and underwater noise pollution, impacts to subsistence hunting) (Arctic Council, 2015; Halliday et al., 2017; Hauser et al., 2018). Commercial shipping mainly uses heavy fuel oil, with associated emissions of sulphur, nitrogen, metals, hydrocarbons, organic compounds and black carbon and fly ash to the atmosphere during combustion (Turner et al., 2017). The use of new technology, like scrubbers, might reduce this impact.

The predominant shipborne activity in Antarctica is fishing, logistic support to land-based stations, and marine research vessels operating for both non-governmental and governmental sectors. Less predictable sea ice conditions and duration pose challenges to these activities (Chown, 2017).

3.3 Polar Ice Sheets and Glaciers: Changes, Consequences and Impacts

3.3.1 Ice Sheet Changes

Over the satellite era, ice sheet mass change has been measured repeatedly using three complimentary satellite methods, and pre–20th century mass changes have been reconstructed using firn/ice core and geological evidence (Appendix 3.A.3.1).

3.3.1.1 West Antarctica and Antarctic Peninsula

Recent studies agree with previous assessments that the West Antarctic Ice Sheet (WAIS) and the Antarctic Peninsula (AP) have lost mass since the early 1990s, that the cumulative loss increased into the 2000s and *very likely* increased further into the last decade in WAIS and *likely* increased further on the AP *(medium evidence, medium agreement)* (Martín-Español et al., 2016; Bamber et al., 2018; Gardner et al., 2018a; Shepherd et al., 2018; Rignot et al., in review) (Figure 3.6, Table 3.3, 3.4).

There is *high agreement* in the sign and *medium agreement* in the magnitude of both WAIS and AP mass change between the three satellite methods for 2003–2010 (Mémin et al., 2015; Shepherd et al., 2018). Disagreements in magnitude between some methods suggest that measurement uncertainties are not fully understood, though all methods now agree with the multi-method mean for WAIS (–93 ± 26 Gt yr^{-1}) and the AP (–27 ± 15 Gt yr^{-1}) (Shepherd et al., 2018).

There is *high confidence* that the rate of WAIS mass loss over the decade since 2007 is greater than over the decade since 1992, reported by two multi–method studies over the same five-year periods and similar spatial extents (Bamber et al., 2018; Shepherd et al., 2018) (Table 3.3), and this acceleration is supported by estimates from separate, overlapping studies (Appendix 3.A.3.1.1) (*robust evidence, medium agreement*).

Table 3.3: Five-year mass balance estimates for WAIS (Shepherd et al., 2018) and WAIS plus a portion of the AP (Bamber et al., 2018).

	1992–1996/7	1997–2001/2	2007–2011/2	2012–2016/7
Mass change (Gt yr^{-1}) (Bamber et al., 2018)	–55 ± 30	–53 ± 30	−197 ± 11	−172 ± 27
Mass change (Gt yr^{-1}) (Shepherd et al., 2018)	–53 ± 29	–41 ± 28	−148 ± 27	−159 ± 26

In agreement with previous assessments, there is *high agreement* that WAIS mass loss and acceleration in loss is concentrated in the Amundsen Sea Embayment (ASE). The main period of ASE loss acceleration occurred in the late 2000s (Mouginot et al., 2014), and losses from 2003–2013 accounted for most of the total WAIS loss of –112 ± 10 Gt yr^{-1} over this period (Martín-Español et al., 2016). The margins of the Getz Ice Shelf also lost mass rapidly (at –67 ± 27 Gt yr^{-1}, 2008–2015) (Gardner et al., 2018a). Regions where the greatest mass loss is currently occurring (such as the ASE) have also experienced mass loss during previous warm periods (Cross-Chapter Box 6). On the AP, mass loss has increased from the 1990s to the last decade (Table 3.4) (Appendix 3.A.3.1.1), with *medium confidence* and *high agreement.*

Table 3.4: Five–year mass balance estimates for the AP (Shepherd et al., 2018).

	1992–1996/7	1997–2001/2	2007–2011/2	2012–2016/7
Mass change (Gt yr^{-1}) (Shepherd et al., 2018)	–7 ± 13	–6 ± 13	–35 ± 17	–33 ± 16

Overall, ongoing, rapid mass loss in the ASE sector of WAIS is *very likely*, with *high confidence* on the magnitude and acceleration of loss from multiple techniques. Total Antarctic Ice Sheet (AIS; combined AP, WAIS and EAIS) mass balance is summarized in Appendix 3.A Table 4.

3.3.1.2 East Antarctica

Changes in East Antarctic Ice Sheet (EAIS) mass assessed in recent studies remain close to zero, with large interannual variability and no clear trend over the satellite record (Table 3.5, Figure 3.6, Appendix 3.A.3.1.2), with *medium evidence* and *high agreement*. The mass signal has an apparent 4.7–year periodicity (Mémin et al., 2015).

The large uncertainties for these measurements result particularly from poorly constrained glacial isostatic adjustment signals in this region (contributing to the changing gravity field), and sparsely observed surface mass balance and firn densification over the extensive EAIS (affecting the satellite altimetry and input-output budgeting methods (Appendix 3.A.3.1)) (Velicogna et al., 2014; Martin-Español et al., 2017; Bamber et al., 2018; Shepherd et al., 2018).

Table 3.5: Five-year mass balance estimates for EAIS (Bamber et al., 2018; Shepherd et al., 2018).

	1992–1996/7	1997–2001/2	2007–2011/2	2012–2016/7
Mass change (Gt yr^{-1}) (Bamber et al., 2018)	+28 ± 76	–50 ± 76	+80 ± 17	–19 ± 20
Mass change (Gt yr^{-1}) (Shepherd et al., 2018)	+11 ± 58	+8 ± 56	+23 ± 38	–28 ± 30

As well as varying temporally, the mass balance of individual EAIS drainage basins also varies spatially from –17 ± 4 Gt yr^{-1} in the Totten/Moscow University/Fox Glacier area (with an acceleration of –4 ± 0.7 Gt yr^{-2}) to +63 ± 6 Gt yr^{-1} (acceleration +15 ± 0.9 Gt yr^{-2}) for 2003–2013, indicating that regions contribute differently (Velicogna et al., 2014). Palaeo ice sheet evidence suggests that sectors of Wilkes Land including Totten Glacier also experienced mass loss in previous warm climate intervals (*limited evidence, high agreement*) (Aitken et al., 2016; Wilson et al., 2018).

In summary, the mass balance of the EAIS is *likely* close to zero, with *medium confidence*, but potentially important glacier changes are *likely* taking place in the Wilkes Land sector, with *high confidence*. Total AIS (combined AP, WAIS and EAIS) mass balance is summarized in Appendix 3.A, Table 4.

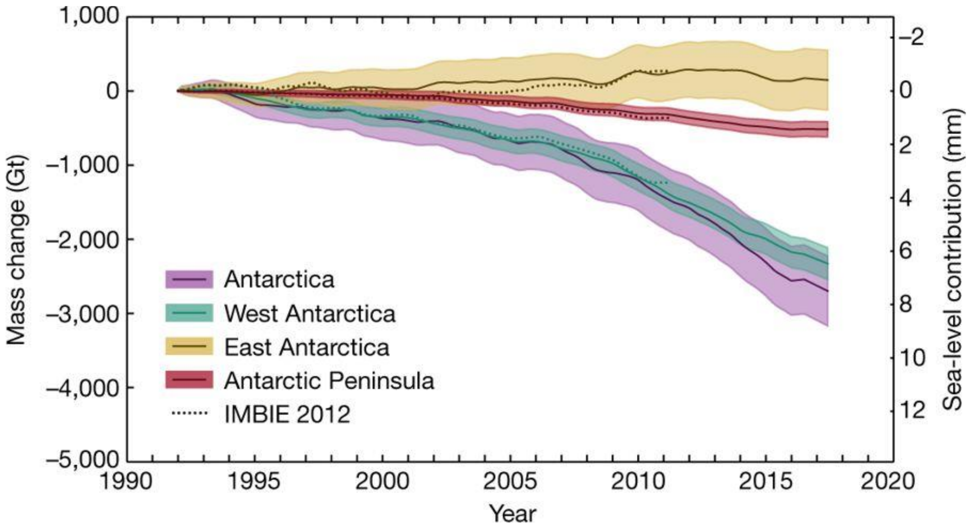


Figure 3.6: Cumulative Antarctic Ice Sheet mass change since 1992. The estimated 1–sigma uncertainty is shaded (Shepherd et al., 2018). [PLACEHOLDER FOR FINAL DRAFT: Figure to be redrafted]

3.3.1.3 Mechanisms of Mass Change: Antarctica

Two mechanisms dominate the signals of mass change observed in Antarctica: changes in surface mass balance (SMB) driven largely by changes in snowfall (not melting), and changes in glacier flow rate that largely control ice discharge from the grounded ice sheet to the sea.

The mass loss trends observed over recent decades from WAIS and the AP are *very likely* dominated by glacier flow acceleration (dynamic thinning) rather than SMB (Appendix 3.A, Figure 8). Loss due to dynamic thinning was –112 ± 10 Gt yr^{-1} for the 2003–2013 period, largely attributable to acceleration of glaciers in the ASE (Appendix 3.A, Figure 8) (Martín-Español et al., 2016), which contributed –102 ± 10 Gt yr^{-1} from 2003–2011 with an acceleration of –15.7 ± 4.0 Gt yr^{-2} (Sutterley et al., 2014). Total ice discharge in the ASE increased by 77% since the 1970s, driven primarily by acceleration of Pine Island Glacier that began around 1945, Smith, Pope and Kohler glaciers around 1980, and Thwaites Glacier around 2000 (Mouginot et al., 2014; Konrad et al., 2017; Smith et al., 2017c). Glacier flow acceleration in the ASE and western AP accounted for 88% of the 36 ± 15 Gt yr^{-1} increase in AIS mass loss from 2008 to 2015 (Gardner et al., 2018a). Ice flow acceleration of up to 25% has also been observed along the Getz Ice Shelf margin between 2007 and 2014 (Chuter et al., 2017).

Dynamic thinning is associated with grounding line retreat, which has been observed with *medium evidence* and *high agreement* in coastal WAIS and on some EAIS and AP glaciers within the satellite era (Rignot et al., 2014; Christie et al., 2016; Hogg et al., 2017; Konrad et al., 2018). A recent study covering most of the Antarctic coast for 2010–2016 found that 22%, 3% and 10% of surveyed grounding lines in WAIS, EAIS and the AP retreated at rates faster than 25 m yr^{-1} (the average pace since the Last Glacial Maximum), with the highest rates (up to 420 m yr^{-1}) along the Amundsen and Bellingshausen Sea coasts of WAIS and the AP, but also on Frost and Totten glaciers of EAIS (up to 200 m yr^{-1}) (Konrad et al., 2018). Retreat rates are variable through time and have previously reached 1–2 km yr^{-1} in the ASE in the 1996–2008 period (Mouginot et al., 2014).

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Dynamic thinning and retreat have been driven primarily by ice-shelf thinning due to basal melting (*medium evidence, high agreement*) which in WAIS increased by 70% in the decade to 2012, and in the ASE averages 8% thickness loss from 1994–2012, greatest near glacier grounding lines (Paolo et al., 2015). ASE ice-shelf basal melting, grounding line retreat and glacier dynamic thinning have varied spatially and temporally with variations in ocean forcing (*limited evidence, medium agreement*) (Paolo et al., 2015; Christianson et al., 2016; Jenkins et al., 2018) with an apparent decadal cycle locally causing a fourfold swing in basal melt rates, which may have dominated glacier mass losses (Jenkins et al., 2018) (*low evidence, low agreement*) or may be superimposed on a sustained and accelerating trend of mass loss compatible with the onset of marine ice sheet instability (*medium evidence, medium agreement*) (Favier et al., 2014; Joughin et al., 2014; Rignot et al., 2014; Christianson et al., 2016) (Cross-Chapter Box 6).

On the AP, large fluctuations in SMB in recent years partially mask a dominant trend of dynamic mass loss (*medium evidence, medium confidence*) (Appendix 3.A, Figure 8) from glaciers in Graham Land (Pritchard et al., 2012; Mouginot et al., 2014; Rignot et al., in review) and the Bellingshausen Sea coast of the southern AP (Wouters et al., 2015; Hogg et al., 2017; Martin-Español et al., 2017), which has gone from being close to balance in the 2000s to -56 ± 8 Gt yr^{-1} loss rate since 2009 (Wouters et al., 2015).

Ice loss driven by glacier acceleration has *likely* continued over recent years for some EAIS drainage basins in Wilkes Land (Flament and Rémy, 2012; Serreze et al., 2016), though SMB fluctuations *very likely* dominate EAIS mass balance on these timescales (Appendix 3.A, Figure 8). Significant short–term regional SMB fluctuations have been observed in Dronning Maud Land (e.g., an anomalous +350 Gt, equivalent to ~1 mm of global sea level drop, for 2009–2011) (Boening et al., 2012; Lenaerts et al., 2013; Welker et al., 2014). On a decadal timescale, however, a study using 76 shallow firn cores from coastal and interior Dronning Maud Land indicate that it is *very likely*, with *medium confidence*, that there is no trend in accumulation between 1950 and 2010 in this sector of the EAIS (Altnau et al., 2015).

Over the past century (1900–2010), firn and ice cores reveal positive accumulation trends in the AP and separate areas of positive, negative and zero trends in WAIS (Thomas et al., 2015; Wang et al., 2017) (*medium confidence*). Antarctica has *likely* experienced a snowfall-driven growth of $+4.3 \pm 1$ Gt per decade during the 19th century increasing to $+14 \pm 1$ Gt per decade during the last 100 years with *medium evidence* (based on 49 ice-core records in EAIS, 7 records in the AP, and 23 in WAIS) (Thomas et al., 2017). EAIS contributed 10% of that growth (+0.8 Gt per decade) on the interior plateau, the Weddell Sea coast and Dronning Maud Land. On the AP, the accumulation increase began in the 1930s and accelerated in the 1990s (Thomas et al., 2015; Goodwin et al., 2016). Increased EAIS SMB mitigated 20th century global sea level rise by 7.7 ± 4.0 mm and WAIS SMB by 2.8 ± 1.7 mm (with *medium evidence* and *medium confidence*), and the more recent SMB increases on the AP indicate that it has begun to mitigate sea level rise by 6.2 ± 1.7 mm per century, based on 53 ice cores records spanning 1801–2000 (Medley and Thomas, Submitted). On longer time scales, four ice cores spanning the last 1000 years suggest an accumulation decrease (Thomas et al., 2015) or, from 67 cores spanning the last 800 years, a statistically negligible change over most of Antarctica, with contemporary SMB not exceptionally high compared to the last 800 years (Frezzotti et al., 2013) (*medium evidence, low confidence*).

Ice sheet basal melting is an additional component of mass balance, not described in AR5. Around 50% of the AIS bed is wet (Siegert et al., 2017), and basal melting produces ~65 Gt yr^{-1} of subglacial water (Pattyn, 2010). This water partly refreezes on the ice sheet sole (Bell, 2008) and partly accumulates in depressions as subglacial lakes, of which over 400 exist beneath the AIS with a total volume of tens of thousands of cubic kilometres (Siegert, 2017), including the largest, subglacial Lake Vostok (around 6000 km^3) (Popov and Masolov, 2007; Lipenkov et al., 2016). Lakes are hydrologically connected by subglacial channels and exist under most of Antarctica's fast-flowing ice streams, and subglacial water flow extends to the grounding line, where it exchanges fresh water and nutrients with the ocean (Section 3.3.3.3) (Fricker et al., 2007; Siegert et al., 2007; Carter and Fricker, 2012; Horgan et al., 2013; Le Brocq, 2013; Flament et al., 2014; Siegert et al., 2016). Subglacial water affects ice dynamics by lubricating glacier sliding and changes in ice-sheet surface slope can divert subglacial drainage, changing ice flow (Fricker et al., 2016). Hydrostatic instability of Antarctic subglacial lakes has been suggested (including for Lake Vostok; Erlingsson (2006)), but further investigation of lake morphology showed that Lake Vostok is *very unlikely* to experience a catastrophic discharge (Richter et al., 2014). In WAIS, *limited evidence* suggests sub-glacial meltwater production is influenced by an active volcanic geothermal heat source (Loose et al., 2018).

Despite *medium agreement* about the importance of subglacial hydrology for ice sheet dynamics, there is *limited evidence* of how the subglacial hydrological system of the polar ice sheets will respond to climate change, and how it may affect ice dynamics and mass balance.

3.3.1.4 Greenland

The Greenland Ice Sheet (GIS) is *virtually certain* to have lost mass since the early 1990s, and currently represents the largest single contributor to ongoing mean sea level rise (van den Broeke et al., 2017). Recent results support previous assessments in showing with *high confidence* a marked shift to strongly and increasingly negative mass balance for the GIS between the early 1990s and mid–2000s, and show that high rates of ice loss have continued (*robust evidence, high agreement)*. A geodetic reconstruction of past ice sheet elevations indicates a GIS mass change of –75.1 ± 29.4 Gt yr^{-1} from 1900 to 1983, –73.8 ± 40.5 Gt yr^{-1} from 1983 to 2003, and –186.4 ± 18.9 Gt yr^{-1} from 2003 to 2010, with the losses consistently concentrated along the west and southeast coasts, though intensifying and spreading to the north and northeast coasts in the latest period (Kjeldsen et al., 2015)

These results are supported by a multi-method satellite assessment (Table 3.6) (Bamber et al., 2018), and by similar results for overlapping periods (Appendix 3.A.3.1.3). Palaeo evidence also suggests that the GIS has contributed to sea level rise during past warm intervals (Cross-Chapter Box 6).

Table 3.6: GIS mass change (Bamber et al., 2018).

	1992–1997	1997–2002	2007–2012	2012–2017
Mass change (Gt yr^{-1})	+31 ± 83	–47 ± 81	–320 ± 10	–247 ± 15

3.3.1.5 Mechanisms of Mass Change: Greenland

Ongoing GIS mass loss over recent years has resulted from a combined increase in surface melting and glacier acceleration. Of these, surface melting now dominates (*medium evidence, high agreement)*, accounting for 42% of losses for 2000–2005, 64% for 2005–2009 and 68% for 2009–2012 (Figure 3.7) (Enderlin et al., 2014; Andersen et al., 2015; Fettweis et al., 2017; van den Broeke et al., 2017). In the early 1990s, the GIS appears to have been close to balance (Hanna et al., 2013; Khan et al., 2015) or gaining mass (+54 ± 48 Gt yr^{-1} for 1961–1990) (Colgan et al., 2015) but a significant summer warming of +2°C since the early 1990s increased modelled GIS surface melt by 35% and runoff by > 40%, with little change in precipitation and sublimation (Hanna et al., 2012; Box, 2013; Van den Broeke et al., 2016; Fettweis et al., 2017).

The post–1990s period experienced the warmest GIS near-surface summer air temperatures of the 1840–2010 period (+1.1˚C) having followed an approximately linear increase that is statistically highly significant, while autumn temperatures changed by +2.1˚C, increasing the length of the melt season (Box, 2013). Runoff totals remain uncertain, however, because although gridded SMB fields from regional climate models (RCMs) agree reasonably well with observations (Lucas-Picher et al., 2012; Fettweis et al., 2013a; Noël et al., 2015), they do not currently resolve the narrow coastal outlet glaciers where melt is greatest, typically leading to an underestimate in RCM surface melt (Noël et al., 2016).

Sub-surface meltwater storage and transport in perennial firn aquifers has recently been observed in south and west GIS (Humphrey et al., 2012; Forster et al., 2013; Kuipers Munneke et al., 2014a) (*medium evidence*), buffering the GIS runoff response (Poinar et al., 2017). These aquifers cover at least 18,000 km^2 at an average elevation of ~1600 m (Miège et al., 2016) and store ~140 Gt of water (Koenig et al., 2014), a volume that can be compared to estimates from RCM output for 1961 to 1990 of total ice sheet melt of 435 Gt yr^{-1} and runoff of 260 Gt yr^{-1} (Van den Broeke et al., 2016). The aquifers have spread to higher altitudes (Steger et al., 2017) and are storing 22% of the increased meltwater production (Noël et al., 2017), but firn densification associated with firn warming of up to +6°C between 1951/2 and 2013 is reducing storage capacity (Polashenski et al., 2014).

The remaining 40% of non-SMB GIS mass loss from 1991 to 2015 has resulted from dynamic mass loss (Flowers, 2018) (Figure 3.7). Since 2000, glacier acceleration accounts for –739 ± 29 Gt, of which 77% occurred on 15 glaciers, with 50% from only four (Enderlin et al., 2014). As these large, tidewater-terminating glaciers have thinned, the dynamic contribution to GIS mass loss has, however, decreased from 58% from 2000 to 2005 to 32% between 2009 and 2012 (Enderlin et al., 2014) (or 17% greater than this if 1961–1990 positive mass balance was part of a long-term trend (Colgan et al., 2015)). Furthermore, it is now apparent that increased surface melt does not lead to sustained increases in glacier flow rate on annual timescales because subglacial drainage networks evolve to cope with the additional water inputs (Sole et al., 2013; Tedstone et al., 2015; Stevens et al., 2016; Nienow et al., 2017).

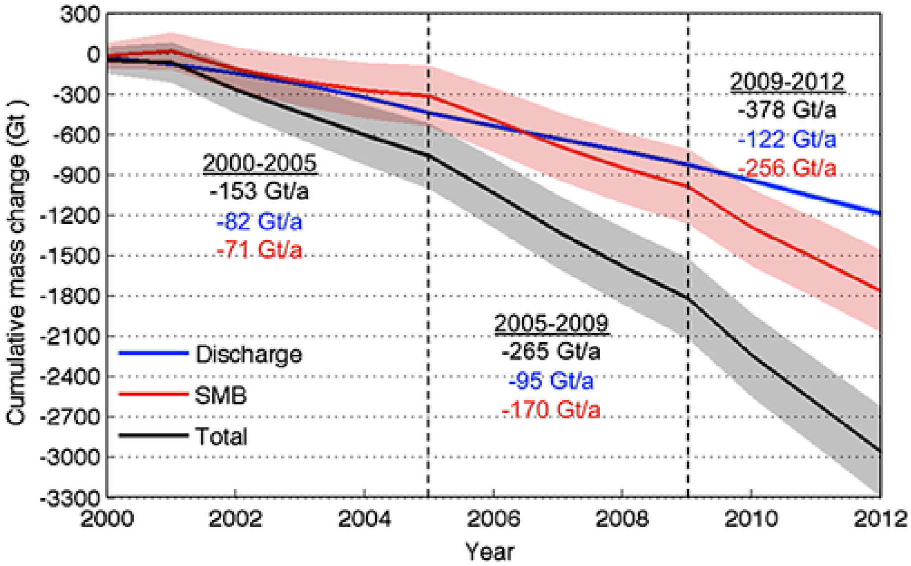

Figure 3.7: GIS cumulative mass change (black) due to discharge change (blue) with respect to 1996 and surface mass balance change (red) with respect to the 1961–1990 mean. Shading indicates uncertainty. Discharge uncertainty is shown but is visually indistinguishable (Enderlin et al., 2014). [PLACEHOLDER FOR FINAL DRAFT: Figure to be redrafted]

3.3.1.6 Drivers

Interlinked changes in the flow of heat from the oceans and in heat and moisture from the atmosphere (Appendix 3.A.1) are the major drivers of observed change in glacier dynamics and SMB on the AIS and GIS, though the behaviour of the ice sheets is subject to feedbacks that can amplify the response to forcing (see Cross-Chapter Box 6).

3.3.1.6.1 Ocean

It is *very likely* that ocean heat forcing is an important driver of dynamic change in the GIS and AIS through the processes of submarine melting and iceberg calving, though several of the processes acting at the ice-ocean interface are poorly observed and not yet well understood.

In Antarctica, it is *likely* that loss of ice-shelf buttressing is the dominant trigger of observed dynamic mass loss, though important feedbacks may significantly affect the rate and extent of ice loss (Section 3.3.1.3 and Cross-Chapter Box 6). Floating ice shelves buttress 90% of AIS outflow either through lateral shear at coastal embayments or vertical shear at pinning points (Depoorter et al., 2013; Rignot et al., 2014; Fürst et al., 2016; Reese et al., 2018), but simulating the impacts of buttressing in ice-sheet models remains a major challenge and research focus (Matsuoka et al., 2015; Pattyn et al., 2017).

Longer and more detailed records of ice-shelf change (Paolo et al., 2015; Christianson et al., 2016; Khazendar et al., 2016) and ocean properties (Jenkins et al., 2016; Webber et al., 2017) have contributed to *high confidence* that the buttressing effect has been reduced by sub-ice-shelf melting and thinning in the ASE, and *medium evidence* for this process outside the ASE (Khazendar et al., 2013; Pollard et al., 2015; Cook et al., 2016; Rintoul et al., 2016; Walker and Gardner, 2017; Adusumilli et al., 2018; Minchew et al., 2018). Around most of the Antarctic coast, near-freezing ocean waters on the continental shelf shield the ice shelves from warmer Circumpolar Deep Water that is found further offshore, but in the Amundsen and Bellingshausen seas, this water intrudes onto the continental shelf, driving sub-ice-shelf melt rates two orders of magnitude higher than elsewhere in Antarctica (Jacobs et al., 1996; Jourdain et al., 2017). There is *limited evidence* that changes in the thickness of the Circumpolar Deep Water layer have controlled recent variability in ice shelf melting (Jacobs et al., 2011; Dutrieux et al., 2014; Christianson et al., 2016; Jenkins et al., 2018).

There is *medium confidence* that changes in winds drive the changes in Circumpolar Deep Water layer thickness. Winds act either directly on cold surface waters within the Ekman layer and warm Circumpolar Deep Water within a shelf-edge undercurrent (Walker et al., 2013; Dutrieux et al., 2014; Kimura et al., 2017), or indirectly through their influence on buoyancy forcing in coastal polynyas which causes deepening of the cold surface layer (St-Laurent et al., 2015; Webber et al., 2017). Winds over the Amundsen Sea are highly variable owing to complex interactions between the SAM, ENSO, Atlantic Multidecadal Oscillation, and the Amundsen Sea Low (Uotila et al., 2013; Li et al., 2014; Turner et al., 2016) (Appendix 3.A.1.3). There is *limited evidence* that ENSO, or other tropical-ocean related variability triggered change on Pine Island Glacier in the 1940s (Smith et al., 2017c) and again in the 1970s and 1990s (Jenkins et al., 2018), while recent changes in ice shelf thickness are correlated with ENSO variability (Paolo et al., 2018). Coupling between wind variability, ocean upwelling, ice shelf melt and glacier flow rate has also been observed at Totten Glacier, Wilkes Land, EAIS (Greene et al., 2017).

Around Greenland, recent studies indicate with *medium confidence* that an anomalous inflow of subtropical water driven by wind changes, multi-decadal natural ocean variability (Andresen et al., 2012), and a long-term increase in the North Atlantic's upper ocean heat content since the 1950s (Cheng et al., 2017), all contributed to a warming of the subpolar North Atlantic (Häkkinen et al., 2013). Modelling studies indicate with *medium confidence* that water temperatures near the grounding zone of GIS outlet glaciers are critically important to their calving rate (O'Leary and Christoffersen, 2013), and warm waters have been observed with *high confidence* interacting with major GIS outlet glaciers (Holland et al., 2008). However, the processes behind warm-water incursions in coastal Greenland, forcing glacier retreat, remain unclear (Straneo et al., 2013; Xu et al., 2013b; Bendtsen et al., 2015; Murray et al., 2015; Cowton et al., 2016; Miles et al., 2016).

Regional-scale ice sheet models have been developed to simulate the advance and retreat of major outlet glaciers (Todd and Christoffersen, 2014; Morlighem et al., 2016; Muresan et al., 2016; Bondzio et al., 2017). Better representation is needed, however, of submarine melt rates and feedbacks on calving rate (Rignot et al., 2010; Todd and Christoffersen, 2014; Benn et al., 2017a) (that are understood with *low confidence*), and the important local characteristics of bed and fjord geometry, ice melange, and subglacial discharge (Enderlin et al., 2013; Gladish et al., 2015; Slater et al., 2015; Morlighem et al., 2016). Overall there is *low confidence* in understanding how coastal GIS outlet glaciers respond to ocean heat forcing, and extrapolation from a small sample of studied glaciers is impractical (Moon et al., 2012; Carr et al., 2013; Straneo et al., 2016; Cowton et al., 2018).

3.3.1.6.2 Atmosphere

Snow accumulation and surface melt in Antarctica are influenced by the Southern Hemisphere extratropical circulation (Appendix 3.A.1.3), which has *likely* intensified and shifted poleward in austral summer between 1950 and 2012 (Arblaster et al., 2014; Swart et al., 2015a). There is *medium confidence* in these changes given the limited observations and spread in magnitude across various datasets and reanalyses. Paleoclimate reconstructions suggest the austral summer SAM has been in its most positive extended state for the past 600 years (Abram et al., 2014; Dätwyler et al., 2017). Intensified atmospheric circulation from 1979 to 2013 has *likely* caused snowfall increases on the western side of the AP and decreases on the eastern side (Marshall et al., 2017) (*medium confidence*). The geographically-variable accumulation trends (1900–2010) across WAIS (Section 3.2.1.2) are explained by a deepening of the prevailing Amundsen Sea Low pressure pattern over recent decades (Raphael et al., 2016) (*high confidence*).

Runoff of surface melt water remains a minor component of AIS mass balance, but melting on ice shelves can cause collapse, abrupt loss of buttressing and glacier acceleration. Extensive drainage networks have recently been mapped that, at least since the mid–20th century, have routed meltwater from the flanks of the AIS down to the ice shelves and, in some cases, to the sea (Bell et al., 2017; Kingslake et al., 2017). Some shelves are *likely* vulnerable to destabilisation if the meltwater supply increases, for example through hydro-fracture, except where efficient surface drainage systems can evolve to drain excess water to the sea (Bell et al., 2017; Kingslake et al., 2017). In coastal EAIS, increased surface melt has been partly caused by katabatic winds (Lenaerts et al., 2016a).

New temperature reconstructions show little change in EAIS surface air temperatures but decadal (1958–2012) warming over WAIS and the AP (Nicolas and Bromwich, 2014). During the 1990s, WAIS experienced record warmth relative to the past 200 years, though ice core records show that similar conditions occurred 1% of the time in the 2000 years prior (Steig et al., 2013). In the AP, summer temperatures are frequently high enough to cause melting. Warming of the northeastern AP began around 600 years ago, and the high rate of warming over the past century is unusual in the context of natural climate variability over the past two millennia (Mulvaney et al., 2012). The melt-related collapse of Larsen B ice shelf, eastern AP, in 2002 is unprecedented over at least the last 11,000 years (Domack et al., 2005) and followed strong warming between the mid–1950s and the late 1990s. Increased föhn winds resulting from a more positive SAM (Cape et al., 2015) have *likely* caused increased melting on the Larsen ice shelves (Grosvenor et al., 2014; Luckman et al., 2014; Elvidge et al., 2015), including during the winter season (Kuipers Munneke et al., 2018a). The AP has cooled since the late 1990s, however, and AP and WAIS decadal temperature changes are now seen as being within the extreme natural variability of the regional atmospheric circulation, and not primarily associated with the drivers of global temperature change (*high agreement, medium evidence*) (Turner et al., 2016; Smith and Polvani, 2017).

Synoptic-scale accumulation and ablation events that may additionally influence the annual AIS mass budget include katabatic winds that sublimate a significant fraction (17%) of snowfall before it reaches the ground (Grazioli et al., 2017), 'atmospheric rivers' causing regional snowfall anomalies such as in Dronning Maud Land (2009–2011) (Gorodetskaya et al., 2014), and the impact of sea ice cover on inland precipitation rates (Lenaerts et al., 2016b) (*low evidence, low confidence*).

In Greenland, associations between atmospheric indices such as the North Atlantic Oscillation (NAO), temperature and snowfall indicate with *high confidence* that, as in Antarctica, variability of large-scale atmospheric circulation is an important driver of GIS SMB changes (Ding et al., 2014; Ding et al., 2017). An increase in negative NAO conditions explain about 70% of summer warming since 2003 (Fettweis et al., 2013b; Mioduszewski et al., 2016) (*medium confidence*) and increased accumulation on the northern/western GIS, caused by enhanced southerly flow of warm, moist air masses into Baffin Bay (Mernild et al., 2015; Osterberg et al., 2015; Wong et al., 2015). An increase in warm air advection associated with a blocking high-pressure system culminated in July 2012 in exceptional warmth and surface melt up to the summit of the ice sheet (Nghiem et al., 2012; Tedesco et al., 2013; Hanna et al., 2014; Hanna et al., 2016; McLeod and Mote, 2016). Reduced snow cover and increased concentrations of organic matter on the ice surface are an additional driver of surface melt in Greenland through a reduction in summer albedo (Box et al., 2012; Tedesco et al., 2016; Stibal et al., 2017) (*medium confidence*).

3.3.1.7 Attribution to Natural and Anthropogenic Forcing

For both Greenland and Antarctica, a longer observational record and improvements in ice sheet modelling are needed before ice sheet mass changes can be attributed with confidence to natural or anthropogenic drivers, particularly in the representation of coupled of atmosphere-ice-ocean processes over appropriate timescales (Section 3.3.1.6), and feedbacks (Cross-Chapter Box 6).

Antarctic SMB is *likely* influenced by trends in atmospheric circulation, but the partitioning between natural and human drivers of circulation changes remains uncertain. Attribution is challenging because atmospheric changes are affected by a combination of greenhouse gas increases, stratospheric ozone depletion (Waugh et al., 2015; England et al., 2016) and considerable natural variability, including in tropical Pacific sea surface conditions (Schneider et al., 2015a; England et al., 2016; Raphael et al., 2016; Clem et al., 2017a) (*medium confidence*). *Limited evidence* from climate modelling suggests that an anthropogenic signal in Antarctic snowfall will emerge from this natural variability by the mid–21st century (Previdi and Polvani, 2016).

Although there is *limited evidence* of an emerging role for anthropogenic forcing of GIS surface melting and interior snowfall trends (Fyke et al., 2014), significant interannual variability in atmospheric circulation currently limits confidence in attributing GIS mass loss acceleration (Wouters et al., 2013) to anthropogenic or natural climate change (*medium confidence*).

3.3.2 Polar Glacier Changes

3.3.2.1 Observations, Mechanisms and Drivers

Here we consider polar glaciers in the Canadian and Russian Arctic, Svalbard, Greenland and Antarctica, independent of the main ice sheets (Figure 3.8). Glaciers in all other regions are assessed in Chapter 2.

It is *very likely* that Arctic glaciers have lost significant mass between 1961 and 2016 (Zemp et al., Submitted), despite considerable inter-annual variations (Figure 3.8). Updated ice mass changes calculated from satellite gravimetry for 2002–2016 are: Arctic Canada –68.1 ± 9.7 Gt yr^{-1}, Russian Arctic –14.5 ± 6.5 Gt yr^{-1}, and Svalbard –9.0 ± 2.8 Gt yr^{-1} (Ciracì et al., In Review). Mass loss of peripheral glaciers in Greenland was –40.9 ± 16.5 Gt yr^{-1} between October 2003 and March 2008, a rate of loss 3–4 times higher per unit area than for the GIS (Bolch et al., 2013). Direct glacier length and mass balance records indicate that the current rate and magnitude of this mass loss is *likely* to be unprecedented for the period of historic observations (Zemp et al., 2015). Pre-historic glacial deposits show that Arctic glaciers may have been smaller than present or may have disappeared altogether in the mid Holocene (Solomina et al., 2015), although this *evidence* is *limited* and does not allow assessment of how these earlier periods of mass loss compare to recently observed changes. Data from ice cores suggest that the current rate of glacier mass loss in Arctic Canada is larger than at any time during the past 4000 years (Fisher et al., 2012; Zdanowicz et al., 2012).

Glacier mass loss in the Arctic has been driven by changes in SMB, glacier dynamics and subaqueous melt, although these mechanisms exhibit spatiotemporal variability and their relative impact is poorly understood. Increased surface melt on Arctic glaciers can lead to a positive feedback from lowered surface albedo, causing further melt (Box et al., 2012), and in Svalbard, there is *limited evidence* that mean glacier albedo has reduced between 1979 and 2015 (Möller and Möller, 2017). Across the Arctic, increased surface melt also reduces the ability of snow and firn to store meltwater, increasing runoff (Zdanowicz et al., 2012; Gascon et al., 2013a; Gascon et al., 2013b; Noël et al., 2017).

Between the 1990s and 2017, it is *likely* that most tidewater glaciers have decelerated in Arctic Canada and accelerated in Svalbard and the Russian Arctic (Strozzi et al., 2017a). Annual retreat rates of tidewater glaciers in Svalbard and the Russian Arctic for 2000–2010, have increased by a factor 2 and 2.5 respectively, between 1992 and 2000 (Carr et al., 2017) (*medium evidence*). Acceleration due to surging (an internal dynamic instability) of a few key glaciers has dominated regional mass loss on time-scales of years to decades (Van Wychen et al., 2013; Dunse et al., 2015).

There is *limited evidence* that calving rates in Svalbard are linked to ocean temperatures which control rates of submarine melt (Luckman et al., 2015). There is also *limited evidence* that the recent acceleration and surge behaviour of polythermal glaciers in Svalbard and the Russian Arctic is caused by destabilization of the marine termini due to increased surface melt, and changes in basal temperature, lubrication and weakening of subglacial sediments (Dunse et al., 2015; Sevestre et al., 2018; Willis et al., 2018) or oceanic forcing and terminus thinning (McMillan et al., 2014a). Rapid disintegration of ice shelves in the Canadian and Russian Arctic continues and has *very likely* led to acceleration and thinning in tributary glacier basins (Willis et al., 2015; Copland and Mueller, 2017).

The main drivers of ice loss from Arctic glaciers are *likely* atmospheric circulation changes. These have led to different rates of retreat between eastern and western glaciers in Greenland's periphery (Bjørk et al., 2018), and have driven increased surface melt in the Canadian Arctic (Gardner et al., 2013; Van Wychen et al., 2015; Millan et al., 2017) through persistently high summer surface temperatures (Bezeau et al., 2014; McLeod and Mote, 2016). The role of anthropogenic forcing in driving these circulation patterns, however, remains unclear (Belleflamme et al., 2015) (*medium confidence*).

On the AP, it is *likely* that independent glaciers have lost mass since the mid–20th century, retreating from positions that they established in the early to mid-Holocene (Ó Cofaigh et al., 2014). Of 860 marine-terminating glaciers, 90% have reduced in area from their earliest recorded positions in the 1940s (Cook et al., 2014). Glaciers in the northeast and southwest AP continue to lose mass in response to ice shelf collapses in the 1970s, 1995 and 2002 and remain far from a state of equilibrium (Scambos et al., 2014; Zhao et al., 2017; Rott et al., 2018; Seehaus et al., 2018). Widespread AP glacier retreat and ice-surface lowering (Fieber et al., 2018) is also consistent with early 21st century satellite-based mass loss estimates that suggest that the (combined) mass loss from glaciers and the ice sheet on the AP is around –30 Gt yr^{-1} (Section 3.3.1.1).

It is *likely* that widespread glacier retreat on the western AP is due to warming of the mid-level ocean, rather than increased air temperature (Wouters et al., 2015; Cook et al., 2016). Small increases in air temperature might in the near future cause a large increase in surface melt for glaciers on the AP (Schannwell et al., 2016; Huber et al., 2017) and in the South Shetland Islands (Falk et al., 2018). On the Kerguelen Islands, *limited evidence* suggests that observed glacier mass loss (Verfaillie et al., 2015) may be dominantly due to migration of storm tracks and associated atmospheric drying rather than increased air temperature (Favier et al., 2016). *Limited evidence* on historic and Holocene changes in Sub-Antarctic glacier change is available (Hodgson et al., 2014), and we have *low confidence* in observed mass changes of independent glaciers across the entire Antarctic and sub-Antarctic region (Figure 3.8).

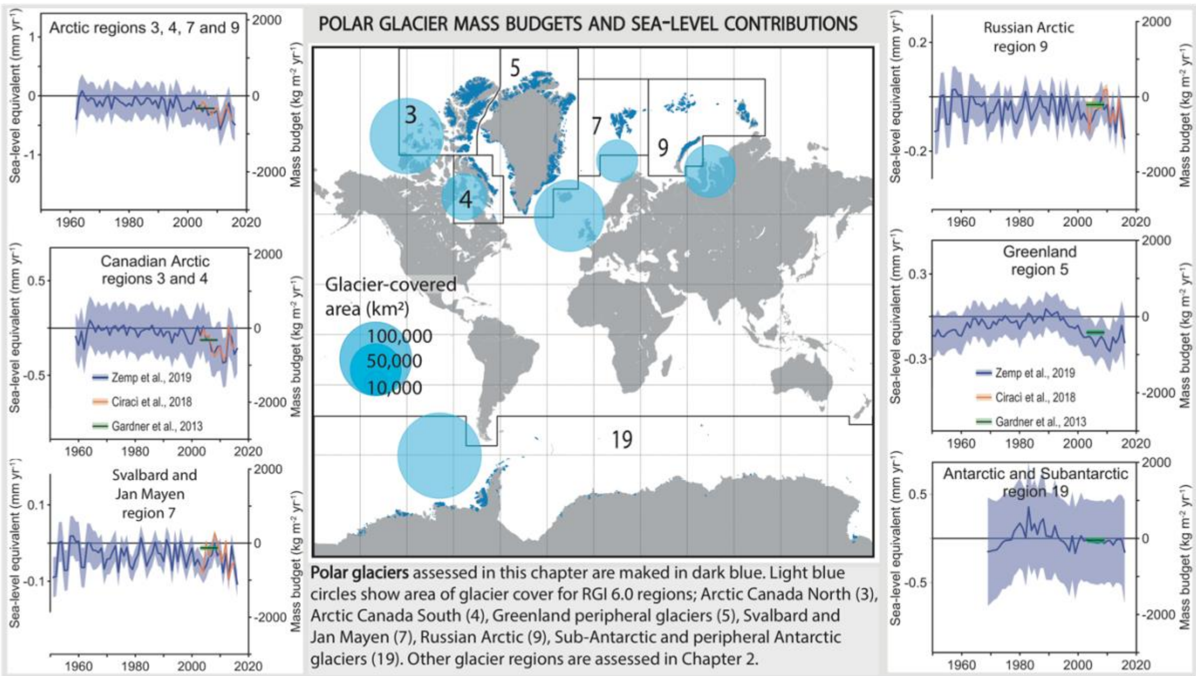

Figure 3.8: Distribution and area of glaciers in selected polar regions (Pfeffer et al., 2014; RGI Consortium, 2017). Mass changes for each region are from (Gardner et al., 2013; Ciracì et al., In Review; Zemp et al., Submitted).

3.3.2.2 Projections

It is *very likely* that glaciers in polar regions will lose substantial mass by the end of the 21st century, with *medium agreement* that losses will be greater under RCP8.5 than RCP2.6 (Radić et al., 2014; Huss and Hock, 2015; Hock et al., Submitted) (Appendix 3.A. Table 5). Although all polar glaciers are projected to lose mass, the extensive glaciers in Arctic Canada and Antarctica are projected to make larger contributions to sea-level rise (Chapter 4) during the 21st century in spite of their smaller projected relative ice losses (Figure 3.8), compared to areas such as Svalbard with higher projected rates but less extensive glaciers (Hock et al., Submitted) (Appendix 3.A. Table 5). We have *medium confidence* in the magnitude and timing of these regional-scale glacier projections because they have been carried out using models that have simplified SMB forcing and ice dynamics.

Though projections have been made for individual polar glaciers using more complete SMB (Lenaerts et al., 2013) or ice-dynamic models (Gilbert et al., 2016; Zekollari et al., 2017), we cannot use these studies to predict the regional behaviour of polar glaciers (Section 3.3.3). Confidence in regional-scale projections will increase when such models include ice dynamics (Clarke et al., 2015; Maussion et al., 2018), subaqueous melt and calving processes (Huss and Hock, 2015; McNabb et al., 2015; Schannwell et al., 2016) and instability mechanisms (Dunse et al., 2015; Sevestre et al., 2018; Willis et al., 2018), and are calibrated against new datasets of observed glacier mass changes (Zemp et al., 2015; Ciracì et al., In Review).

3.3.3 Consequences and Impacts

3.3.3.1 Sea Level

Chapter 4 assesses the sea level impacts from observed and projected changes in ice sheets (Section 3.3.1) and polar glaciers (Section 3.3.2), including uncertainties related to marine ice sheets (Cross-Chapter Box 6).

3.3.3.2 Physical Oceanography

The major large-scale impacts of freshwater discharge from Greenland on ocean circulation relate to the potential modulation/inhibition of the formation of water masses that represent the headwaters of the Atlantic Meridional Overturning Circulation. The timescales and likelihood of such effects are assessed separately in Chapter 6.

For Antarctica, freshwater input to the ocean from the ice sheet is divided approximately equally between melting of calved icebergs and of ice shelves in situ (Depoorter et al., 2013; Rignot et al., 2014). There is *high confidence* that the input of ice shelf meltwater has increased in the Amundsen and Bellingshausen Seas since the 1990s, but *low confidence* on trends in other sectors (Paolo et al., 2015).

Freshwater injected from the AIS affect water mass circulation and transformation, though sea ice dominates upper ocean properties away from the Antarctic ice shelves (Abernathey et al., 2016; Haumann et al., 2016). Over the ice-shelf regions, where dense waters sink and flood the global ocean abyss, the role of glacial freshwater input is clearer. From 1980–2012, the salinity of Antarctic Bottom Water reduced by an amount equivalent to 73 ± 26 Gty^{-1} of freshwater added, around half the estimated increase in freshwater input by Antarctic glacial discharge up to that time (Purkey and Johnson, 2013). In some places, notably the Indian-Australian sector, Antarctic Bottom Water freshening may be accelerating (Menezes et al., 2017). There is *medium confidence* in an overall freshening trend and *low confidence* that this is accelerating, given *limited evidence* and significant interannual variability in Antarctic Bottom Water properties at other export locations (Meijers et al., 2016).

For the Southern Ocean, there is *limited evidence* for stratification changes in the post-AR5 period, and a variety of potential mechanisms for change. An increase in stratification caused by strengthened discharge of freshwater from the AIS was invoked as a mechanism to suppress vertical heat flux and permit an increase in sea ice extent (Bintanja et al., 2013), though some studies conclude that glacial freshwater input is insufficient to cause a significant sea ice expansion (Swart and Fyfe, 2013; Pauling et al., 2017) (see also Section 3.2.1.1). In contrast, where warm water intrusions drive melting within ice shelf cavities, a significant entrained heat flux to the surface can exist and increase stratification and potentially reduce sea ice extent (Jourdain et al., 2017; Merino et al., 2018). It has been argued that freshening from glacial melt can enhance basal melting of ice shelves by reducing dense water production and modulating oceanic heat flow into ice-shelf cavities (Silvano et al., 2018).

3.3.3.3 Biogeochemistry

The ice sheets in both polar regions have the potential to deliver significant amounts of nutrients and organic carbon to the ocean, including via subglacial meltwater, icebergs, surface runoff and melting of the base of ice shelves (Shadwick et al., 2013; Wadham et al., 2013; Hood et al., 2015; Raiswell et al., 2016; Yager et al., 2016; Herraiz-Borreguero et al., 2016b; Hodson et al., 2017) (Figure 3.9). This may stimulate primary production during the summer melt season (Bhatia et al., 2013; Hawkings et al., 2015; Hawkings et al., 2016; Wadham et al., 2016) (*medium confidence*), and may increase carbon drawdown from the atmosphere.

Future predictions of these processes are made more challenging by the landward progression of marine-terminating glaciers and collapse of ice shelves (Cook et al., 2016). This has the potential to drive major shifts in nutrient supply to coastal waters (Figure 3.9). *Limited evidence* suggests that erosion of newly-exposed glacial sediments in front of retreating land-terminating glaciers (Monien et al., 2017) and increased diffuse nutrient fluxes from newly exposed glacial sediments on the seafloor (Wehrmann et al., 2014) will amplify nutrient supply, whilst other nutrient sources may be cut off (e.g., icebergs, upwelling of marine water (Meire et al., 2017)).

Observations from a fjord in southwest Greenland indicate that summer phytoplankton blooms are associated with peak meltwater discharge from the ice sheet, and account for a significant proportion of the annual primary production (Juul-Pedersen et al., 2015; Meire et al., 2016a). However, direct measurements suggest that glacial meltwater is a significant source of silica, phosphorous and iron (Hawkings et al., 2015) but not nitrogen, which may ultimately limit the integrated primary production during summer (Meire et al., 2016b; Hopwood et al., 2018). There is *limited evidence* of an increase in dissolved nutrient fluxes from the GIS during high melt years, but the response of the dominant sediment-bound fraction may not increase with rising melt (Hawkings et al., 2015). Thus, there is *low confidence* overall in the magnitude of the response of nutrient fluxes from ice sheets to enhanced melting.

Subglacial discharge plumes, in addition to the direct supply of nutrients contained in the meltwater, may drive an indirect supply of nutrients by entraining nutrient-rich seawater (Meire et al., 2017; Hopwood et al., 2018; Kanna et al., 2018) (*medium evidence*). There is *limited evidence* that upwelled nutrient fluxes may enhance primary production over a distance of 10–100 km along the trajectory of the outflowing plume (Hopwood et al., 2018). There is *high agreement based* on *medium evidence* that long-term tidewater glacier retreat into shallower water or onto land, a plausible scenario for about 55% of the 243 distinct outlet glaciers in Greenland (Morlighem et al., 2017), will reduce or diminish this indirect mechanism of nutrient supply through glacier runoff, thereby reducing summer productivity in Greenland fjord ecosystems (Meire et al., 2017; Hopwood et al., 2018).

For Antarctica**,** there is *high agreement* based on *medium evidence* that enhanced input of iron from ice shelves and glacial meltwater in addition to icebergs can stimulate primary production in polynyas, coastal regions and the wider Southern Ocean (Gerringa et al., 2012; Herraiz-Borreguero et al., 2016b). Findings from Herraiz-Borreguero et al. (2016a) for the Amery Ice Shelf in East Antarctica indicate that marine ice (accreted at the base of the ice shelf) can act as a reservoir and transport pathway for subglacial and/or sediment-derived iron to coastal marine ecosystems. Calving of the Mertz Glacier Tongue in East Antarctica in 2010 led to an input of meltwater which enhanced the availability of light and iron in the Mertz polynya, supporting a diatom bloom that doubled carbon uptake relative to pre-calving conditions (Shadwick et al., 2013) (*low confidence*). Glacial melt and sea ice melt in the Amundsen Sea polynya support strong productivity and high levels of carbon export; this carbon sink could intensify in the short term with increased glacial melt (*low confidence*), but its long-term trajectory is unknown (Yager et al., 2016). Satellite observations and modelling also indicate the potential for icebergs to fertilise the Southern Ocean beyond the coastal zone (Death et al., 2014; Duprat et al., 2016).

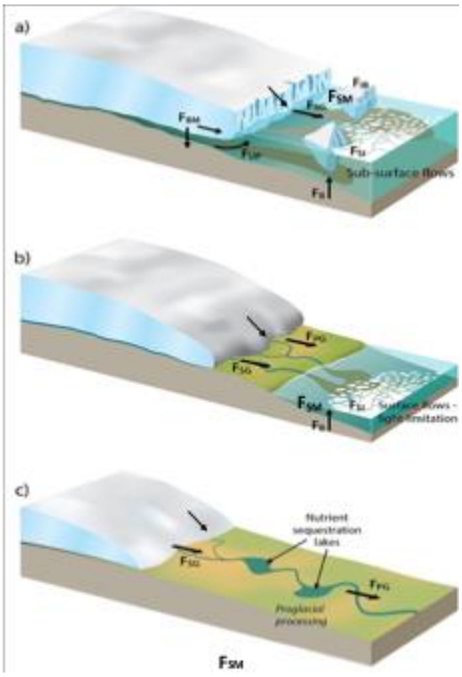

Figure 3.9. Potential shifts in nutrient fluxes (F) with landward retreat of marine-terminating glaciers (a) at different stages (b and c) (BM= basal melting, SG/SM=subglacial/surface melt, IB =icebergs, SI=sea ice, B=benthic (sea-floor), PG=proglacial).

3.3.3.4 Ecosystems

For Greenland and Svalbard there is *high agreement* based upon *limited evidence* that the retreat of marine-terminating glaciers will alter food supply to higher trophic levels of marine food webs (Meire et al., 2017; Milner et al., 2017). The consequences of changes in glacial systems on marine ecosystems are often mediated via the fjordic environments that fringe the edge of the ice sheets, for example changing physical-chemical conditions have affected the benthic ecosystems of Arctic fjords (Bourgeois et al., 2016) (m*edium confidence*). The amplification of nutrient fluxes caused by enhanced upwelling at calving fronts (Meire et al., 2017), combined with high carbon/nutrient burial and recycling rates (Wehrmann et al., 2013; Smith et al., 2015), plays an important role in sustaining high productivity of the Arctic fjord ecosystems of Greenland and Svalbard (Lydersen et al., 2014). Glacier retreat, causing glaciers to shift from being marine-terminating to land-terminating, can reduce the productivity in coastal areas off Greenland with potentially large ecological implications, also negatively affecting production of commercially harvested fish (Meire et al., 2017).

For Antarctica**,** there is *high agreement* based on *medium evidence* that ice-shelf retreat or collapse is leading to new marine habitats and to biological colonization (Gutt et al., 2011; Fillinger et al., 2013; Trathan et al., 2013; Hauquier et al., 2016). The loss of ice shelves and retreat of coastal glaciers around the AP in the last 50 years has exposed at least 2.4×10^4 km^2 of new open water. These newly revealed habitats have allowed new phytoplankton blooms to be produced resulting in new marine zooplankton and seabed communities (Gutt et al., 2011; Fillinger et al., 2013; Trathan et al., 2013; Hauquier et al., 2016), and have resulted in enhanced carbon uptake by coastal marine ecosystems (*medium confidence*), although quantitative estimates of carbon uptake are highly variable (Trathan et al., 2013; Barnes et al., 2018). New available habitat on coastlines may also provide breeding or haul-out sites for land-based predators such as penguins and seals (Trathan et al., 2013) (*low confidence*). Fjords that have been studied in the subpolar western AP are hotspots of benthic abundance and biodiversity (Grange and Smith, 2013) and there is evidence that glacier retreat in these environments can impact the structure and function of benthic communities (Moon et al., 2015; Sahade et al., 2015) (*low confidence*).

[START CROSS-CHAPTER BOX 6 HERE]

Cross-Chapter Box 6: Future Sea Level Changes from Marine Ice Sheets [TBC]

Authors: Rob De Conto (USA), Alexey Ekaykin (Russian Federation), Andrew Mackintosh (New Zealand), Roderik van de Wal (Netherlands), Jeremy Bassis (USA)

Over the last century, glaciers were the main contributors to increasing ocean water mass (Chapter 4), due to their relatively fast response to changing climate (Chapter 2). However, most terrestrial frozen water is stored in Antarctic and Greenland ice sheets (Chapter 3), and future changes in their dynamics and mass balance will cause sea level rise over the 21st century and beyond (Chapter 4).

Greenland's present-day mass balance is dominated by surface melt and runoff (Chapter 3), and its contribution to 21st century sea-level rise will also be dominated by these processes (Chapter 4). In contrast, about a third of the much larger Antarctic Ice Sheet (AIS) rests on bedrock hundreds of meters below sea level (Figure 4.5), with most of the ice-sheet margin terminating directly in the ocean. These features make the overlying ice sheet vulnerable to dynamical instabilities with the potential to cause rapid ice loss (see below).

In many places around the AIS margin, the seaward-flowing ice forms floating ice shelves (Figure CB6.1). Ice shelves in contact with bathymetric features on the sea floor or confined within embayments provide back stress (buttressing) that impedes the seaward flow of the upstream ice and thereby stabilizes the ice sheet. The ice shelves are thus a key factor controlling AIS dynamics. Almost all Antarctic ice shelves provide substantial buttressing (Fürst et al., 2016) but some are currently thinning at an increasing rate (Khazendar et al., 2016). Today, thinning and retreat of ice shelves is associated primarily with ocean-driven basal melt that, in turn, promotes iceberg calving (see Section 3.3.2.3).

Accumulation and percolation of surface melt and rain water also impact ice shelves by lowering albedo, deepening surface crevasses, and causing flexural stresses that can lead to hydrofracturing and ice shelf collapse (Macayeal and Sergienko, 2013). In some cases supraglacial rivers might diminish destabilizing impact of surface melt by removing meltwater before it ponds on the ice-shelf surface (Bell et al., 2017). In summary, both ocean forcing and surface melt affect ice shelf mechanical stability (*high confidence*), but the precise importance of the different mechanisms remains poorly understood and observed.

Future response of the AIS to warming will largely be determined by changes in ice shelves, because their thinning or collapse will reduce their buttressing capacity, leading to an acceleration of the grounded ice and to thinning of the ice margin. In turn, this thinning can initiate grounding line retreat (Konrad et al., 2018). If the grounding line is located on bedrock sloping downwards toward the ice sheet interior (retrograde slope), initial retreat can trigger a positive feedback, due to non-linear response of the seaward ice flow to the grounding line thickness change. As a result, progressively more ice will flow into the ocean (Figure CB6.1). This self-sustaining process is known as Marine Ice Sheet Instability (MISI) (AR5). The onset and persistence of MISI is dependent on several factors in addition to overall bed slope, including the details of the bed geometry and conditions, ice-shelf pinning points, lateral shear from the walls, self-gravitation effects on local sea level and isostatic adjustment. Hence, long-term retreat on every retrograde-sloped bed is not necessarily unstoppable (Gomez et al., 2015).

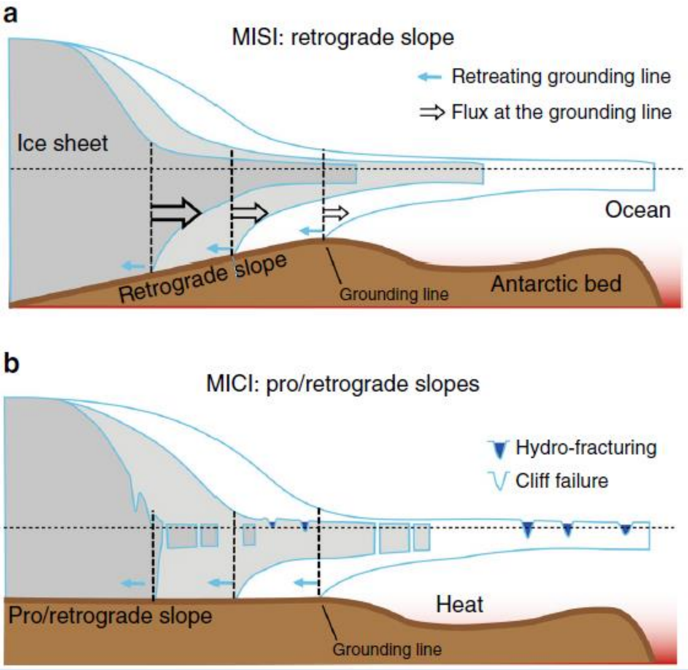

Figure CB6.1: Schematic representation of Marine Ice Sheet Instability (MISI, a) and Marine Ice Cliff Instability (MICI, b) from Pattyn (2018). a - thinning of the buttressing ice shelf leads to acceleration of the ice sheet flow and thinning of the marine-terminated ice margin. Because bedrock under the ice sheet is sloping towards ice sheet interior, thinning of the ice causes retreat of the grounding line followed by an increase of the seaward ice flux, further thinning of the ice margin, and further retreat of the grounding line. b - disintegration of the ice shelf due to bottom melting and/or hydro-fracturing produces an ice cliff. If the cliff is tall enough (~ 800 m of total ice thickness, or about 100 m of ice above the water line), the stresses at the cliff face exceed the strength of the ice, and the cliff fails structurally in repeated calving events. Note that MISI requires a retrograde bed slope, while MICI can be realized on a flat or seaward inclined bed. Like MISI, the persistence of MICI depends on the lack of ice-shelf buttressing, which can stop or slow brittle ice failure at the grounding line by providing supportive backstress.

The MISI process might be particularly important in West Antarctica, where most of the ice sheet is grounded on bedrock below sea level (Figure 4.5). Since AR5, there is growing observational and modelling evidence that accelerated retreat may be underway in several major Amundsen Sea outlets, including Thwaites, Pine Island, Smith, and Kohler glaciers (e.g., Rignot et al., 2014) supporting the MISI hypothesis, although observed grounding-line retreat on retrograde slope is not definitive proof that MISI is underway.

It has recently been shown (Barletta et al., 2018) that the Amundsen Sea Embayment experiences unexpectedly fast bedrock uplift (41 mm per year) as an adjustment to reduced ice mass loading, which could help stabilize grounding line retreat.

One of the largest outlets of the East Antarctic Ice Sheet, Totten glacier, has also been retreating and thinning in recent decades (Li et al., 2015). Totten’s current behaviour suggests that East Antarctica could become a substantial contributor to future sea level rise, as it has been in the past (Aitken et al., 2016). It is not clear, however, if the recently observed changes are a linear response to increased ocean forcing, or an indication that MISI has commenced (Roberts et al., 2017).

Changes in ice dynamics are thought to be less important for Greenland Ice Sheet (GIS) compared to AIS. The GIS has limited direct access to the ocean, through relatively narrow subglacial channels (Morlighem et al., 2014), and most of the bedrock at the ice-sheet margin is above sea level (Figure 4.5). However, since AR5 it has been shown that several Greenland outlet glaciers (Petermann, Kangerdlugssuaq, Jakobshavn Isbræ, Helheim, Zachariæ Isstrøm) and North-East Greenland Ice Stream may contribute more than expected to future sea level rise (Mouginot et al., 2015). It has also been shown Greenland was nearly ice free for extended periods during the Pleistocene, suggesting a sensitivity to deglaciation under climates similar to or slightly warmer than present (Schaefer et al., 2016).

The disappearance of ice shelves allows formation of ice cliffs, which may be inherently unstable if they are tall enough to generate stresses that exceed the strength of the ice. This ice cliff failure can lead to ice sheet retreat via a process called marine ice cliff instability (MICI; Figure CB2.1), that has been hypothesized to cause partial collapse of the West Antarctic Ice Sheet within a few centuries (Pollard et al., 2015; DeConto and Pollard, 2016).

Limited evidence is available to confirm the importance of MICI. In Antarctica, marine-terminating ice margins with the grounding lines thick enough to produce unstable ice cliffs are currently buttressed by ice shelves, with a possible exception of Crane glacier on the Antarctic Peninsula (AR5). In Greenland, MICI-style behavior is seen today at the termini of Jakobshavn and Helheim glaciers (James et al., 2014), but calving of these narrow outlets is controlled by a combination of ductile and brittle processes, which might not be representative examples of much wider Antarctic outlet glaciers, like Thwaites.

Overall, there is *low agreement* on the exact MICI mechanism and *low evidence* of its occurrence in the present or the past. Thus the potential of MICI to impact the future sea level remains very uncertain.

Limited evidence from paleo-environmental records suggests that parts of AIS experienced rapid (i.e., on centennial time-scale) retreat *likely* due to ice sheet instability processes since Last Glacial Maximum. Geological records and ice sheet modelling suggests that the AIS experienced multiple periods of rapid ice loss between 20,000 and 9,000 years ago and particularly during Melt Water Pulse –1a, a centennial-scale global sea level rise ~ 14,600 years ago (Golledge et al., 2014; Weber et al., 2014). Both the WAIS (including Pine Island glacier) and the EAIS also experienced rapid thinning and grounding line retreat during the early to mid-Holocene (Jones et al., 2015b; Wise et al., 2017). In the Ross Sea, grounding lines may have retreated several hundred kilometers inland and then re-advanced to their present-day positions due to bedrock uplift after ice mass removal (Kingslake et al., 2018), thus supporting the stabilizing role of glacial isostatic adjustment on ice sheets (Barletta et al., 2018). These past rapid changes have *likely* been driven by the incursion of Circumpolar Deep Water onto the Antarctic continental shelf (Golledge et al., 2014; Hillenbrand et al., 2017) and MISI (Jones et al., 2015b). *Limited* geological *evidence* of past MICI in Antarctica is provided by deep iceberg plough marks on the sea-floor (Wise et al., 2017).

The ability of models to simulate the processes controlling MISI has improved since AR5 (Pattyn, 2018), but significant discrepancies in projections remain (see Chapter 4), due to poor understanding of mechanisms, and due to lack of the observational data to constrain the models. Inclusion of MICI in one ice sheet model has improved its ability to match albeit uncertain geological sea level targets in the Pliocene (Pollard et al., 2015) and Last Interglacial (DeConto and Pollard, 2016), although the MICI solution may not be unique (Aitken et al., 2016) (see Chapter 4).

Overall, we conclude that rapid retreat and thinning of some Antarctic and to a lesser extent Greenland outlet glaciers is underway, with significant implications for future sea level (Chapter 4). However, the timescale and future rate of these processes is not well known, casting deep uncertainty on ice sheet and sea level projections.

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[END CROSS-CHAPTER BOX 6 HERE]

3.4 Terrestrial Cryosphere: Changes, Consequences and Impacts

3.4.1 Observations

3.4.1.1 Seasonal Snow Cover

Terrestrial snow cover is a defining characteristic of the Arctic land surface for up to 9 months each year, with changes influencing the surface energy budget, ground thermal regime, and freshwater budget. Snow cover also interacts with vegetation, influences biogeochemical activity, and affects habitats and species, with consequences for ecosystem services. Arctic land areas north of 60°N are always completely snow covered in winter, so the transition seasons of autumn and spring are key when characterizing variability and change.

3.4.1.1.1 Extent and duration

Dramatic reductions in Arctic spring snow cover extent have occurred since satellite charting began in 1967 (Estilow et al., 2015), with declines in May and June of –3.1% and –13.6% per decade (relative to the 1981–2010 mean (Derksen et al., 2017); updated through 2018) (Figure 3.10) (*high confidence*). Trends calculated from independent data sources covering shorter time periods are consistent with the 1967–2018 data for May, but there is large inter-dataset spread for June (–5% to –14% per decade) (Hori et al., 2017; Mudryk et al., 2017).

The loss of spring snow extent is reflected in shorter snow cover duration derived from surface observations (Bulygina et al., 2011; Brown et al., 2017a), satellite data (Wang et al., 2013; Estilow et al., 2015), and model-based analyses (Liston and Hiemstra, 2011) (*high confidence*). These trends range between –0.7 and –3.9 days per decade depending on region and time period, but all spring snow cover duration trends from all datasets are negative (Brown et al., 2017a). These same multi-source datasets also identify reductions in autumn snow extent and duration (–0.6 to –1.4 days per decade; summarized in (Brown et al., 2017a) (*high confidence*). While positive trends in October and November SCE are apparent in a single dataset (Hernández-Henríquez et al., 2015), they are not replicated in other surface, satellite, and model datasets (Brown and Derksen, 2013; Mudryk et al., 2017).

3.4.1.1.2 Depth and water equivalent

Weather station observations across the Russian Arctic identify negative trends in the maximum snow depth before melt between 1966 and 2014 (Bulygina et al., 2011; Osokin and Sosnovsky, 2014). There is *medium confidence* in these observations because the pointwise nature of these measurements do not capture prevailing conditions across the landscape. Seasonal maximum snow depth trends over the North American Arctic are mixed and largely statistically insignificant (Vincent et al., 2015; Brown et al., 2017a). The timing of maximum snow depth has shifted earlier by 2.7 days per decade for the North American Arctic (Brown et al., 2017a); comparable analysis is not available for Eurasia.

Gridded products from remote sensing and land surface models identify negative trends in snow water equivalent between 1981 and 2016 for both the Eurasian and North American sectors of the Arctic (Brown et al., 2017a). While the snow water equivalent anomaly time series show reasonable consistency between products when averaged at the continental scale, considerable inter-dataset variability in the spatial patterns of change (Liston and Hiemstra, 2011; Park et al., 2012; Brown et al., 2017a) mean there is only *medium confidence* in these trends.

3.4.1.1.3 Drivers

Despite uncertainties due to sparse observations (Cowtan and Way, 2014), surface temperature has increased across Arctic land areas in recent decades (Hawkins and Sutton, 2012; Fyfe et al., 2013), driving reductions in Arctic snow extent and duration (Section 3.4.1.1.1) (*very high confidence).* Seasonal warming maxima in the autumn and spring periods (Brown et al., 2017a) are consistent with observed delays to the start of the snow accumulation season, a reduced fraction of precipitation falling as snow (Screen and Simmonds, 2011), and earlier melt in the spring (Brown and Derksen, 2013). Changes in Arctic snow extent can be directly related to extratropical temperature increases (Brutel-Vuilmet et al., 2013; Thackeray et al., 2016; Mudryk et al., 2017). Based on multiple historical datasets, there is a consistent temperature sensitivity for Arctic snow extent, with approximately 800 000 km^2 of snow cover lost per °C warming in spring (Brown and Derksen, 2013; Brown et al., 2017a), and 700 000 to 800 000 km^2 lost in autumn (Derksen and Brown, 2012; Brown and Derksen, 2013).

Snowfall-based drivers of Arctic snow cover changes are highly uncertain because precipitation remains a sparse and highly uncertain measurement over Arctic land areas. While efforts are underway to improve the coordinated correction of systematic precipitation measurement errors (Kochendorfer et al., 2017), in situ datasets remain uncertain (Yang, 2014) and largely regional (Kononova, 2012; Vincent et al., 2015). Atmospheric reanalyses provide another perspective on Arctic precipitation (Vihma et al., 2016) but these products are inconsistent and poorly validated (Serreze et al., 2012). Previous assessments have identified positive trends in Arctic precipitation (Min et al., 2008; Callaghan et al., 2011; Hartmann et al., 2013), but more recent assessments are not available.

Arctic snow accumulates to the height of the prevailing ground vegetation after which it is redistributed by wind to topographic depressions and drifts (Sturm and Stuefer, 2013). Despite improved process understanding, estimates of sublimation loss during blowing snow events remain a key uncertainty in the mass budget of the Arctic snowpack. Increased shrub cover influence snow capture and soil temperatures (Gouttevin et al., 2012; Druel et al., 2017), but changes in vegetation cover across the Arctic (at the coherent regional-scales needed to impose an impact on the hydro-climatic system) are not uniform and the drivers are poorly understood (Myers-Smith et al., 2015) (see Section 3.4.3.2.1). Vegetation changes can also influence spring snow melt via changes to albedo (Marsh et al., 2010; Loranty et al., 2014).

Darkening of snow through the deposition of black carbon (BC) and other light absorbing impurities (Bullard et al., 2016) enhances snow melt (Hansen and Nazarenko, 2004) (*very high confidence*). The global direct radiative forcing for BC in seasonal snow and sea ice is estimated to be 0.04 W m^{-2}, but the effective forcing can be up to threefold greater at regional scales due to the enhanced albedo feedback triggered by the initial darkening (Bond et al., 2013). Lawrence et al. (2011) estimate the present-day radiative effect of BC and dust in land-based snow to be 0.083 W m^{-2}, only marginally greater than the simulated 1850 effect (0.075 W m^{-2}) due to offsetting effects from increased BC emissions and reductions in dust darkening and snow cover (*medium confidence*). Kylling et al. (2018) estimate a surface radiative effect of 0.292 W m^{-2} caused by dust deposition (largely transported from Asia) to Arctic snow, approximately half of the BC central scenario estimate of Flanner et al. (2007). The forcing from brown carbon deposited in snow (associated with both combustion and secondary organic carbon) is estimated to be 0.09−0.25 W m^{-2}, with the range due to assumptions of particle absorptivity (Lin et al., 2014) (*low confidence*).

The influence of changing Arctic sea ice conditions on seasonal terrestrial snow is an emerging area of research. Reanalyses and model simulations suggest increasing atmospheric moisture in the Arctic in response to reduced sea ice extent (Liu et al., 2012; Screen et al., 2013; Petrie et al., 2015). Temperature and snowfall responses over Eurasia have been statistically associated with regions of sea ice loss (Mori et al., 2014; Wegmann et al., 2015) but the circulation impacts and driving mechanisms remain uncertain (Li and Wu, 2012; Barnes and Screen, 2015).

3.4.1.2 Frozen Ground

3.4.1.2.1 Temperature

Record high temperatures at ~10–20 m depth in the permafrost (below the depths affected by intra-annual fluctuation in temperature) have been documented at most long-term monitoring sites in the Northern Hemisphere circumpolar permafrost zone (AMAP, 2017) (Figure 3.10) (*high confidence*). At some locations, the temperature is 2°C–3°C higher than 30 years ago. Since 2000, the typical rate of increase in permafrost temperatures was between 0.4°C and 0.7°C per decade for colder continuous permafrost monitoring sites and between 0.1°C and 0.2°C for warmer discontinuous permafrost. Relatively smaller increases in permafrost temperature in warmer sites indicate that near-surface permafrost is thawing with additional heat absorbed by the ice-to-water phase change, and as a result, the active layer is increasing in thickness. In contrast to temperature, there is only *medium confidence* that active layer thickness has been observed to increase. This is because decadal trends vary across regions and sites (Shiklomanov et al., 2012) and because mechanical probing of the active layer can underestimate the degradation of near-surface permafrost in some cases because the surface subsides when ground ice melts and drains (Streletskiy et al., 2017). Site averages in three of six Arctic study regions (Russian Far East, Russian European North, East Siberia) show a decadal trend of increasing active layer thickness, whereas three other regions (West Siberia, North Slope Alaska, Northwest Canada) do not show this trend (Romanovsky et al., 2016; AMAP, 2017). Permafrost in the Southern Hemisphere polar region occurs in ice-free exposed rock areas, 0.18% of the total land area of Antarctica (Burton-Johnson et al., 2016). This area is three orders of magnitude smaller than the 13–18x10^6 km^2 area underlain by permafrost in the Northern Hemisphere permafrost zone (Gruber, 2012). Antarctic permafrost temperatures are generally colder and increasing in 3 of 4 measurement locations (Noetzli et al., 2017).

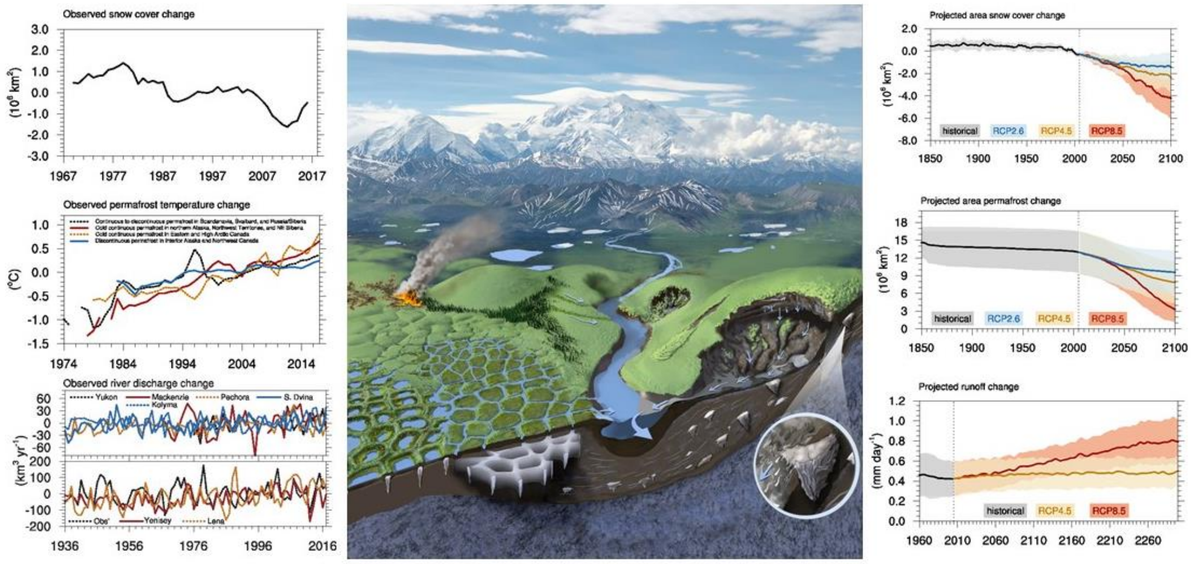


Figure 3.10: [Note for review: This draft graphic will be standardized for final production; high level comments at this stage are helpful, but many smaller points are already undergoing change]. Schematic of important land surface processes influenced by the Arctic terrestrial cryosphere. Left column: time series of snow cover extent anomalies in May (from the NOAA snow chart data record; relative to 1981–2010 climatology; Estilow et al. (2015)), permafrost temperature change normalized to a baseline period (Romanovsky et al., 2017b) and runoff from northern flowing watersheds normalized to a baseline period (1981–2010) (www.arcticgreatrivers.org). Right column: projected future changes to in May snow cover extent and area of near-surface permafrost from the CMIP5 multi-model average for different Representative Concentration Pathway scenarios, and runoff projected from eight Earth System models that contributed to the Permafrost Carbon Network Model Intercomparison experiment (McGuire et al., 2018; Andresen et al., Submitted).

3.4.1.2.2 Ground ice

Permafrost thaw and loss of ground ice causes the land surface to subside and collapse into the volume previously occupied by ice, resulting in disturbance to overlying ecosystems and human infrastructure (Jorgenson et al., 2013; Raynolds et al., 2014). Excess ice in permafrost is typical, ranging for example from 40% of total volume in some sands up to 80–90% of total volume in fine-grained soil/sediments (Kanevskiy et al., 2013). Ice-rich permafrost where impacts are greatest includes the Yedoma deposits in Siberia, Alaska, and the Yukon in Canada, with ice roughly divided between massive wedges interspersed with frozen soil/sediment containing pore ice and smaller ice features (Zimov et al., 2006; Schirrmeister et al., 2011; Strauss et al., 2017). Other regions including Northwestern Canada, the Yamal and Gydan peninsulas of West Siberia, and smaller portions of Eastern Siberia and Alaska contain buried glacial ice bodies of significant thickness and extent (Lantuit and Pollard, 2008; Leibman et al., 2011; Kokelj et al., 2017). Higher resolution ground ice maps exist in some regions where economic development resulted in engineering geological assessments of permafrost for planning purposes (Stephani et al., 2014; Trochim et al., 2016; Vincent et al., 2017), but this resolution is still lacking at the pan-Arctic scale (Jorgenson and Grosse, 2016) even though new remote-sensing technologies are being developed (Minsley et al., 2012).

3.4.1.2.3 Carbon

The permafrost zone represents a large, climate-sensitive reservoir of organic carbon with the potential for some of this pool to be rapidly decayed and transferred to the atmosphere as carbon dioxide or methane as permafrost thaws in a warming climate, thus accelerating the pace of climate change (Schuur et al., 2015). The current best estimate of total (surface plus deep) organic soil carbon (terrestrial) in the northern circumpolar permafrost zone (17.8×10^6 km^2 area) is 1460 to 1600 petagrams (*medium confidence*) (Pg; 1 Pg = 1 billion metric tons) (Schuur et al., 2018). This inventory includes all soil orders within the permafrost zone and thus also counts carbon in nonpermafrost soil orders, active layer (surface) carbon that thaws seasonally, and peatlands. All permafrost-zone soils estimated to 3 m in depth (surface) contain 1035 ± 150 Pg C (Hugelius et al., 2014) (*high confidence*), with two-thirds of the soil carbon pool in Eurasia, and the remaining one-third in North America (including Greenland) (Tarnocai et al., 2009). Of this amount, 800–1000 Pg C is perennially frozen, with the remainder contained in seasonally thawed soils. The 1035 Pg of soil carbon quantified from the northern circumpolar permafrost zone adds another 50% to the global 3-m inventory (2050 Pg C, excluding tundra and boreal biomes (Jobbágy and Jackson, 2000), even though it occupies only 15% of the total global soil area (Schuur et al., 2015)).

Substantial permafrost carbon exists below 3 m depth (*medium confidence*). Deep carbon (>3m) has been best quantified for the Yedoma region of Siberia and Alaska, characterized by wind- and water-moved permafrost sediments tens of meters thick. The yedoma region covers a $1.4x10^6$ km^2 area that remained ice-free during the last Ice Age (Strauss et al., 2013). The carbon inventory of this region comprises yedoma soils that were previously thawed as lakes formed and then refrozen into permafrost when lakes drained, interspersed by intact permafrost yedoma deposits that were unaffected by thaw-lake cycles (Walter Anthony et al., 2014). Together, this region accounts for 327 to 466 Pg C in deep sediment accumulations below 3 m (Strauss et al., 2017).

The current inventory has also highlighted additional carbon pools that are likely to be present but are so poorly quantified (*low confidence*) that they cannot yet be added into the number reported above. There are deep terrestrial soil/sediment deposits outside of the yedoma region that may contain about 400 Pg C (Schuur et al., 2015). An additional pool is organic carbon remaining in permafrost but that is now submerged on shallow Arctic sea shelves that were formerly exposed as terrestrial ecosystems during the Last Glacial Maximum ~20,000 years ago (Walter et al., 2007). This permafrost is slowly degrading due to seawater intrusion, and it is not clear what amounts of permafrost and organic carbon still remain in the sediment versus what has already been converted to greenhouse gases. A recent synthesis of permafrost extent for the Beaufort Sea shelf showed that most remaining subsea permafrost in that region exists near shore with much reduced area (*high confidence*) as compared to original subsea permafrost maps that outlined the entire $3x10^6$ km^2 shelf area (<120 m sea level depth) that was formerly exposed as land (Ruppel et al., 2016). These observations are supported by modelling that suggests that submarine permafrost would be already thawed >10 m depth or more under current submerged conditions (Anisimov et al., 2012; AMAP, 2017).

3.4.1.2.4 Drivers

Changes in temperature and precipitation act as gradual ‘press’ (i.e., continuous) disturbances that directly affect permafrost by modifying the ground thermal regime, as recorded in the observations of the permafrost borehole network (Biskaborn et al., 2015) and discussed in 3.4.1.3.1. Climate changes also can modify the occurrence and magnitude of abrupt physical disturbances such as fire, and soil subsidence and erosion resulting from ice-rich permafrost thaw (thermokarst). These ‘pulse’ (i.e., discrete) disturbances (sensu, Smith et al., 2009) often are part of the ongoing disturbance and successional cycle in Arctic and boreal ecosystems (Grosse et al., 2011), but changing rates of occurrence alter the landscape distribution of successional ecosystem states, with permafrost characteristics defined by the ecosystem and climate state (Jorgenson et al., 2013).

Pulse disturbances often rapidly remove the insulating soil organic layer, leading to permafrost degradation. Of all pulse disturbance types, wildfire affects the most high-latitude land area annually at the continental scale. There is *high confidence* that area burned, fire frequency, and extreme fire years are higher now than the first half of the last century, or even the last 10,000 years (Kasischke and Turetsky, 2006; Flannigan et al., 2009; Kelly et al., 2013). There is *high confidence* that recent climate warming has been linked to increased wildfire activity in the boreal forest regions in Alaska and western Canada where this has been studied. Based on satellite imagery, an estimated 80,000 km^2 of boreal area was burned globally per year from 1997 to 2011 (van der Werf et al., 2010; Giglio et al., 2013). Extreme fire years in northwest Canada during 2014 and Alaska during 2015 doubled the long-term (1997–2011) average area burned annually in this region (Canadian Forest Service, 2017), surpassing Eurasia to contribute 60% of the global boreal area burned (van der Werf et al., 2010; Randerson et al., 2012; Giglio et al., 2013). These extreme North American fire years were balanced by lower-than-average area burned in Eurasian forests, resulting in a 5% overall increase in global boreal area burned. There is *very high confidence* that changes in the fire regime are degrading permafrost faster than had occurred over the historic successional cycle (Rupp et al., 2016), and that the effect of this driver of permafrost change is underrepresented in the permafrost temperature observation network.

Abrupt permafrost thaw occurs when warming interacts with geomorphological processes. Melting ground ice causes the ground surface to subside, which alters surface hydrology. Pooling or flowing water causes localized permafrost thaw and even mass erosion depending on geomorphological conditions. Together, these localized feedbacks can thaw through meters of permafrost within a short time, much more rapidly than would be caused by increasing air temperature alone. This is a pulse disturbance to permafrost that can occur in response to climate, such as an extreme precipitation event (Balser et al., 2014; Kokelj et al., 2015), or coupled with other disturbances such as wildfire that affects the ground thermal regime (Jones et al., 2015a). There is *low confidence* in the importance of abrupt thaw for driving change in permafrost at the circumpolar scale because it occurs at point locations rather than continuously across the landscape, but the risk for widespread change from this mechanism remains high because of the rapidity of change in these locations. New research at the global scale has revealed that 3.6×10^6 km^2, about 20% of the northern permafrost zone, appears to be vulnerable to abrupt thaw (Olefeldt et al., 2016). This susceptible area contained 31% of the total organic carbon pool stored in the 0–3 m soil and up to 50% of the total carbon pool that includes the deep carbon >3 m, highlighting spatial correlations between processes and features that lead to abrupt thaw and storage of organic carbon.

3.4.1.3 Freshwater Systems

There is increasing awareness of the influence of a changing climate on freshwater systems across the Arctic, and associated impacts on hydrological, biogeophysical, and ecological processes (Prowse et al., 2015; Walvoord and Kurylyk, 2016), and northern populations (Takakura, 2018) (Section 3.4.3.3.1). Assessing these impacts requires consideration of complex inter-connected processes, many of which are incompletely observed. The increasing imprint of human development, such as flow regulation on major northerly flowing rivers adds complexity to the determination of climate-driven changes.

3.4.1.3.1 Freshwater ice

Long-term in situ river ice records indicate that the duration of ice cover in Russian Arctic rivers decreased by 7 to 20 days between 1955 and 2012 (Shiklomanov and Lammers, 2014) (*high confidence*). This is consistent with reductions in Arctic river ice cover derived from models (Park et al., 2015) and regional analysis of satellite data (Cooley and Pavelsky, 2016).

Analysis of satellite imagery between 2000 and 2013 identified a significant trend of earlier spring ice break-up across all regions of the Arctic (Šmejkalová et al., 2016); independent satellite data showed approximately 80% of Arctic lakes have experienced declines in ice cover duration during 2002–2015, due to both a later freeze-up and earlier break-up (Du et al., 2017) (*high confidence*). There are indications that lake ice across Alaska has thinned in recent decades (Alexeev et al., 2016), but ice thickness trends are not available at the pan-Arctic scale. Analysis of satellite data over northern Alaska show that approximately one-third of bedfast lakes (the entire water volume freezes by the end of winter) experienced a regime change to floating ice over the 1992–2011 period (Surdu et al., 2014; Arp et al., 2015). This can result in degradation of underlying permafrost (Arp et al., 2016; Bartsch et al., 2017). Lakes of the central and eastern Canadian High Arctic are transitioning from a perennial to seasonal ice regime (Surdu et al., 2016).

3.4.1.3.2 Surface water and runoff

A high lake fraction is present across ice-rich Arctic land areas because permafrost limits surface water drainage and supports ponding even across areas with high moisture deficits (Grosse et al., 2013). While thaw in continuous permafrost is linked to intensified thermokarst activity and subsequent ponding (resulting in lake/wetland expansion), observations of change in surface water coverage across the Arctic are regionally variable (Nitze et al., 2017; Pastick et al., 2018). In ice-rich regions, degrading polygon landscapes with associated subsidence can reduce inundation, increase runoff, and decrease surface water (Liljedahl et al., 2016; Perreault et al., 2017). In discontinuous permafrost, thaw opens up pathways of subsurface flow, improving the connection among inland water systems which supports the drainage of lakes and overall reduction in surface water cover (Jepsen et al., 2013).

Thermokarst lake expansion has been observed in the continuous permafrost of northern Siberia (Smith et al., 2005; Sannikov, 2012; Polishchuk et al., 2015). Net surface water area reduction has been observed in discontinuous permafrost of central and southern Siberia (Smith et al., 2005; Kirpotin et al., 2008; Sharonov, 2012), Canada (Labrecque et al., 2009; Carroll et al., 2011; Lantz and Turner, 2015) and interior Alaska (Chen et al., 2012; Rover et al., 2012). A loss of pond abundance and coverage has occurred across the Arctic coastal plain of Alaska where permafrost is continuous (Andresen and Lougheed, 2015). Increased evaporation from warmer/longer summers, decreased recharge due to reductions in snow melt volume, and dynamic processes such as ice-jam flooding (Chen et al., 2012; Bouchard et al., 2013; Jepsen et al., 2015) are important considerations for understanding observed surface water area change across the Arctic.

A general trend of increasing discharge has been observed for large Siberian (Peterson et al., 2002; Troy et al., 2012; Walvoord and Kurylyk, 2016) and Canadian (Ge et al., 2013; Déry et al., 2016) rivers that drain to the Arctic Ocean (Figure 3.10) (*medium confidence*). Extreme regional runoff events have also been identified (Stuefer et al., 2017). Between 1976 and 2015, trends are 3.1 ± 2.0% for Eurasian rivers and 2.6 ± 1.7% for North American rivers (Holmes et al., 2015) (Figure 3.10). An observed increase in baseflow in the North American (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009) and Eurasian Arctic (Smith et al., 2007a; Duan et al., 2017) over the last several decades is attributable to permafrost thaw and the concomitant enhancement in groundwater discharge. Increases in baseflow represent a notable heat flux to the Arctic Ocean (Yang et al., 2014). The timing of spring season peak flow is generally earlier (Ge et al., 2013; Holmes et al., 2015). There is consistent evidence of decreasing summer season discharge for the Yenisei, Lena, and Ob watersheds in Siberia (Ye et al., 2003; Yang et al., 2004a; Yang et al., 2004b) and the majority of northern Canadian rivers (Déry et al., 2016). Long-term records indicate water temperature increases (Webb et al., 2008; Yang and Peterson, 2017); attribution to rising air temperatures is complicated by the influence of reservoir regulation (Liu et al., 2005; Lammers et al., 2007).

3.4.1.3.3 Drivers

Increases in poleward atmospheric moisture transport in a warmer Arctic lower troposphere (and hence precipitation; see Box 3.1) are consistent with observed increases in discharge from northern rivers into the Arctic Ocean (Zhang et al., 2013). While a number of products suggest increases in Arctic precipitation in recent decades (Lique et al., 2016; Vihma et al., 2016), there remains *low confidence* in reanalysis-based closure of the Arctic freshwater budget due to a wide spread between available reanalysis derived precipitation estimates (Lindsay et al., 2014). While reductions in summer Arctic snowfall have been identified (Screen and Simmonds, 2011), there is no evidence of trends in rain-on-snow events, which can have important ecological implications (Cohen et al., 2015; Dolant et al., 2017).

Studies suggest increases in satellite and model-derived estimates of evapotranspiration across the Arctic (Rawlins et al., 2010; Liu et al., 2014; Liu et al., 2015a; Fujiwara et al., 2016; Suzuki et al., 2018) (*medium confidence)*. Landscape alterations, including disturbance and shifting vegetation patterns also play a key role in driving changes to freshwater systems (Bring et al., 2016; Wrona et al., 2016). In permafrost regions, surface elevation changes due to thaw subsidence in thermokarst-affected landscapes substantially drive hydrologic change by forming depressions for lake formation or generating lake drainage pathways (Jones et al., 2011; Grosse et al., 2013). Increases in the seasonal active layer thickness impact temporary water storage and thus runoff regimes in drainage basins. Formation of taliks underneath lakes and rivers may result in reconnection of surface with sub-permafrost ground water aquifers with varying hydrological consequences depending on local geological and hydraulic settings (Wellman et al., 2013). Vegetation changes influence hydrological and biogeochemical processes by impacting ground temperatures and permafrost (Nauta et al., 2014).

3.4.2 Projections

3.4.2.1 Seasonal Snow

Historical CMIP5 simulations tend to underestimate reductions in observed spring snow cover extent due to uncertainty in the parameterization of snow processes (Essery, 2013; Thackeray et al., 2014), challenges in simulating snow-albedo feedback (Qu and Hall, 2014; Fletcher et al., 2015; Li et al., 2016), unrealistic temperature sensitivity (Brutel-Vuilmet et al., 2013; Mudryk et al., 2017), and biases in climatological spring snow cover (Thackeray et al., 2016). The role of precipitation biases is not well understood (Thackeray et al., 2016).

Reductions in Arctic snow cover duration are projected by the CMIP5 multi-model ensemble due to later snow onset in the autumn and earlier snow melt in spring (Brown et al., 2017a) driven by increased surface temperature over essentially all Arctic land areas (Hartmann et al., 2013). There is *very high confidence* that projected snow cover declines are proportional to the amount of future warming in each model realization (Thackeray et al., 2016; Mudryk et al., 2017). Projections to mid-century are primarily dependent on natural variability and model dependent uncertainties rather than the choice of forcing scenario (Hodson et al., 2013). By end of century, however, differences between scenarios emerge. RCP4.5 stabilizes at 5–10% Arctic snow cover duration reductions (compared to a 1986–2005 reference period); under RCP8.5, this decline continues to unabated to –15 to –25% (Brown et al., 2017a) (Figure 3.10) (*high confidence*).

Positive Arctic snow water equivalent changes emerge across the eastern Eurasian Arctic by mid-century for both RCP4.5 and 8.5 (Brown et al., 2017a) (*high confidence*). Projected snow water equivalent increases across the North American Arctic are less extensive, and emerge later in the century only under RCP8.5 (Brown et al., 2017a). These projected increases are due to enhanced snowfall (Krasting et al., 2013) from a more moisture rich Arctic atmosphere coupled with temperatures between November and April that remain sufficiently low for precipitation to fall as snow. This is not the case for May through October, and for more temperate regions of the Arctic (i.e., Scandinavia) where temperatures do not remain sufficiently low and precipitation phase changes to rainfall result in projected decreases in snow water equivalent (de Vries et al., 2014; Brown et al., 2017a). Snow water equivalent across large portions of the Arctic is presently unaffected by temperature variability (solely driven by precipitation availability) but this area is projected to decrease by mid-century as temperature forcing of precipitation phase becomes more important (Sospedra-Alfonso and Merryfield, 2017).

Changes in snow properties such as density and stratigraphy, which are highly relevant for understanding the impacts of changes to Arctic snow on ecosystems, cannot be resolved directly by climate model simulations, rather they require detailed snow physics models driven by projected climate forcing.

3.4.2.2 Permafrost

Models at the circumpolar or global scale represent permafrost degradation in response to warming scenarios as increases in active layer thickness only. The CMIP5 models project with *high confidence* that active layers will increase and areal extent of near-surface permafrost will decrease substantially (Koven et al., 2013; Slater and Lawrence, 2013) (Figure 3.10). However, there is only *medium confidence* in the magnitude of these changes due to at least a five-fold range of estimated present day near-surface permafrost area (<5 – $>25x10^6$ km^2) by these models. This was caused by wide range of model sensitivity in permafrost area to air temperature change, resulting in a large range of projected near-surface permafrost loss by 2100: 15–87% under RCP4.5 and 30–99% under RCP8.5. The high warming scenario (RCP8.5) would leave most of the current discontinuous permafrost zone free of near-surface permafrost with the remaining near-surface permafrost located around the coldest regions in the northern hemisphere: northern Siberia and the islands of northeast Canada. A more recent analysis of near-surface permafrost trends from a subset of models that self-identified as structurally representing the permafrost zone had a significantly smaller range of estimated present day near-surface permafrost area ($13.1–19.3x10^6$ km^2) (McGuire et al., 2016). This subset also showed large reductions of near-surface permafrost area under RCP8.5, averaging a 91% loss ($12.7x10^6$ km^2) of permafrost area by 2300, with much of that long-term loss already occurring by 2100 (McGuire et al., 2018).

Pulse disturbances are not included in the permafrost projections described above, and there is *high confidence* that fire and abrupt thaw will accelerate change in permafrost relative to climate effects alone, if the rates of these disturbances increased. The observed trend of increasing fire is projected to continue for the rest of the century across most of the tundra and boreal region for many climate scenarios, with the boreal region projected to have the greatest increase in total area burned (Balshi et al., 2009; Rupp et al., 2016). Due to vegetation-climate interactions, there is only *medium confidence* in projections of future area burned. As fire activity increases, flammable vegetation, such as the black spruce forest that dominates boreal Alaska, is projected to decline as it is replaced by low-flammability deciduous forest (Johnstone et al., 2011). In other regions such as western Canada, by contrast, black spruce could be replaced by the even more flammable jack pine, creating regional-scale feedbacks that increases the spread of fire on the landscape. In tundra regions, graminoid (grass-like) tundra is projected to be replaced by more-flammable shrub tundra in future climate scenarios, and tree migration into tundra could further increase fuel loading (Rupp et al., 2016). In contrast to fire, there is only an initial comprehensive circumpolar projection of how abrupt thaw rates may change in the future, with projections of abrupt thaw area expected to increase three-fold from 0.68×10^6 km^2 to 2.5×10^6 km^2 by 2300 under RCP8.5 (Turetsky et al., Submitted). As a result, there is *low confidence* in the ability to assess risk, even though this mechanism for rapid change appears critically important for projecting future change (Kokelj et al., 2017).

3.4.2.3 Freshwater Systems

Climate model simulations project a warmer and wetter Arctic (Krasting et al., 2013). Specific humidity is projected to increase due to enhanced evaporation (Laîné et al., 2014), and moisture flux convergence increases into the Arctic (Skific and Francis, 2013). Relative humidity increases will be driven by contrasts in heating over land versus ocean, and the influence of this heating on marine air masses advected over land (Vihma et al., 2016).

Increased cold-season precipitation is projected across the Arctic by CMIP5 models (Lique et al., 2016) due to increased moisture flux convergence from outside the Arctic (Zhang et al., 2012) and enhanced moisture availability from reduced sea ice cover (Bintanja and Selten, 2014) (*high confidence*). Increases in precipitation extremes are also projected over northern watersheds (Kharin et al., 2013; Sillmann et al., 2013), while rain on snow events are expected to increase (Hansen et al., 2014). Although evapotranspiration will be enhanced in a warmer Arctic (Laîné et al., 2014) a net increased ratio of precipitation minus evaporation is projected, resulting in increased freshwater flux from the land surface to the Arctic Ocean, projected to be 30% above current values by 2100 under RCP4.5 (Haine et al., 2015) (Figure 3.10). This is consistent with projections of increased discharge from Arctic watersheds (van Vliet et al., 2013). The water temperature of this increased discharge is projected to be approximately 1°C warmer than current conditions, increasing the heat flux to Arctic Ocean (van Vliet et al., 2013). The influence of changing vegetation (Pearson et al., 2013) and permafrost conditions (McGuire et al., 2016) are *likely* to introduce regional variability in the hydrological response to a wetter Arctic.

When forced by regional climate models, lake ice models project an earlier spring break-up of between 10–25 days by mid-century (compared with 1961–1990), and up to a 15-day delay in the freeze-up for lakes in the North American Arctic (Brown and Duguay, 2011; Dibike et al., 2011; Prowse et al., 2011b) (*medium confidence*). More extreme reductions are projected for coastal regions. Mean maximum ice thickness is projected to decrease by 10–50 cm over the same period (Brown and Duguay, 2011). High-latitude warming is projected to drive earlier river ice break-up in spring due to both decreasing ice strength, and earlier onset of peak discharge (Cooley and Pavelsky, 2016). Complex interplay between hydrology and hydraulics in controlling spring flooding and ice jam events (which can be important events for sediment and nutrient transport; Turcotte et al. (2011) reduce confidence in related projections (Prowse et al., 2010; Prowse et al., 2011b).

3.4.3 Consequences and Impacts

3.4.3.1 Global Climate Feedbacks

3.4.3.1.1 Carbon cycle

Climate warming is expected to change the storage of carbon in vegetation and soils in northern regions, and net carbon transferred to the atmosphere as carbon dioxide or methane acts as a feedback to accelerate global climate change. There is *high confidence* that the northern region acted as net carbon sink as carbon accumulated in terrestrial ecosystems over the Holocene (Loisel et al., 2014; Lindgren et al., 2018). There is increasing, but divergent evidence, that changing climate in the modern period has shifted these ecosystem into net carbon sources (*low confidence*). Syntheses of ecosystem CO_2 fluxes have alternately showed tundra ecosystems as carbon sinks or neutral averaged across the circumpolar region for the 1990s and 2000s (McGuire et al., 2012), or carbon sources over the same time period (Belshe et al., 2013). Both syntheses agree that the summer growing season is a period of net carbon uptake into terrestrial ecosystems (*high confidence*), and this uptake appears to be increasing as a function of vegetation density/biomass (Ueyama et al., 2013). The discrepancy between these syntheses may be a result of CO_2 release rates during the non-summer season that are now thought to be higher than previously estimated (*high confidence*) (Natali et al., Submitted) or the separation of upland and wetland ecosystems types that can differ in carbon sink/source strength with wetlands more often than not still acting as annual net carbon sinks (Lund et al., 2010). Recent aircraft measurements of atmospheric CO_2 concentrations over Alaska showed that tundra regions of Alaska were a consistent net CO_2 source to the atmosphere, whereas boreal forest regions were either neutral or net

CO_2 sinks for the period 2012 to 2014 (Commane et al., 2017). That study region as a whole was estimated to be a net carbon source of 25 ± 14 Tg C per year averaged over the land area of both biomes for the entire study period. For comparison to projected global emissions, this would be equivalent to a net source of 0.3 Pg CO_2 per year assuming the Alaska study region (1.6 x 10^6 km^2) could be scaled to the entire northern circumpolar permafrost zone soil area (17.8 x 10^6 km^2).

The permafrost soil carbon pool is climate sensitive and an order of magnitude larger than carbon stored in plant biomass (Schuur et al., in review) (*high confidence*). Initial estimates were converging on a range of cumulative emissions from soils to the atmosphere, but recent studies have actually widened that range somewhat (Figure 3.11) (*medium confidence*). Expert assessment and lab soil incubation studies suggest that substantial quantities of C (tens to hundreds Pg C) could potentially be transferred from the permafrost carbon pool into the atmosphere under a warming climate (RCP8.5) (Schuur et al., 2013; Schädel et al., 2014). Global dynamical models supported these findings, showing potential carbon release from the permafrost zone ranging from 37 to 174 Pg C by 2100 under the RCP8.5 climate warming trajectory, with an average across models of 92 ± 17 Pg C (mean ± SE) (Zhuang et al., 2006; Koven et al., 2011; Schaefer et al., 2011; MacDougall et al., 2012; Burke et al., 2013; Schaphoff et al., 2013; Schneider von Deimling et al., 2015). This range is generally consistent with several newer data-driven modeling approaches that estimated that soil carbon releases by 2100 (for RCP8.5) will be 57 Pg C (Koven et al., 2015) and 87 Pg C (Schneider von Deimling et al., 2015), as well as an updated estimate of 102 Pg C from one of the previous models (MacDougall and Knutti, 2016). However, the latest model runs performed with either structural enhancements to better represent permafrost carbon dynamics (Burke et al., 2017a), or common environmental input data (McGuire et al., 2016) show similar soil carbon losses, but also indicate the potential for stimulated plant growth (nutrients, temperature/growing season length, CO_2 fertilization) to offset some or all of these losses by sequestering new carbon into plant biomass and increasing carbon inputs into the surface soil (McGuire et al., 2018). Overall, the estimates support the idea that the northern permafrost zone could emit carbon on the order similar to other current biospheric sources like land use change, but will generally be only a fraction of fossil fuel emissions (*high confidence*). Furthermore, there is *high confidence* that climate scenarios that involve mitigation (e.g RCP4.5) will help to dampen the response of carbon emissions from the Arctic and boreal regions.

Northern ecosystems contribute significantly to the global methane budget, but there is *low confidence* about the degree to which additional methane from northern lakes, ponds, wetland ecosystems, and the shallow Arctic Ocean shelves is already contributing to increasing atmospheric concentrations. Long-term direct observations of methane dynamics are scarce, and analyses of atmospheric concentrations in Alaska concluded that local ecosystems surrounding the observation site have not changed in the exchange of methane from the 1980s until the present, which suggests that either the local wetland ecosystems are responding in step with other northern wetland ecosystems, or that increasing atmospheric methane concentrations in northern observation sites is derived from methane coming from mid latitudes (Sweeney et al., 2016). But this contrasts with indirect integrated estimates of methane emissions from observations of expanding permafrost thaw lakes that suggest a release of an additional 1.6–5 Tg CH_4 yr^{-1} over the last 60 years (Walter Anthony et al., 2014). At the same time, there is *high confidence* that methane fluxes at the ecosystem to regional scale have been under-observed, in part due to the low solubility of methane in water leading to ebullution (bubbling) flux to the atmosphere that is heterogeneous in time and space. This is reflected in new quantifications of: cold-season methane emissions that can be >50% of the annual budget of terrestrial ecosystems (Zona et al., 2016); geological methane seeps that may be climate sensitive if permafrost currently serves as a cap preventing atmospheric release (Walter Anthony et al., 2012; Ruppel and Kessler, 2016; Kohnert et al., 2017); estimates of shallow Arctic Ocean shelf methane emissions where the range of estimates has increased with more observations and now ranges from very high (17 Tg CH_4 yr^{-1}) (Shakhova et al., 2013), to very low (3 Tg CH_4 yr^{-1}) (Thornton et al., 2016). Observations such as these underlie the fact that source estimates for methane made from atmospheric observations are typically lower than methane source estimates made from upscaling of ground observations (e.g., Berchet et al., 2016), and this problem has not improved, even at the global scale, over several decades of research (Saunois et al., 2016; Crill and Thornton, 2017).

In many of the model projections previously discussed, methane release is not explicitly represented because fluxes are small even though higher global warming potential of methane makes these emissions relatively more important than on a mass basis alone. Global models that do include methane show that emissions may already (from 2000–2012) be increasing at a rate of 1.2 Tg CH_4 yr^{-1} in the northern region as a direct response to temperature (Riley et al., 2011; Gao et al., 2013; Poulter et al., 2017). Future methane emissions are also sensitive to changes in ecosystem water balance, including the extent of high-latitude wetlands that are, in part, a result of restricted water infiltration by permafrost. A modelling intercomparison study forecast northern methane emissions to increase from 18 Tg CH_4 yr^{-1} to 42 Tg CH_4 yr^{-1} under RCP8.5 by the 2100 largely as a result of an increase in wetland extent (Zhang et al., 2017). But, projected methane emissions are sensitive to changes in surface hydrology (Lawrence et al., 2015) and a suite of models that were thought to perform well in high-latitude ecosystems showed a general soil drying trend even as the overall water cycle intensified (Andresen et al., Submitted). Furthermore, most models described above do not include many of the abrupt thaw processes that can result in lake expansion, wetland formation, and massive erosion and exposure to decomposition of previously frozen carbon-rich permafrost, leading to *low confidence* in future model projections of methane. A recent study that does include these processes suggests that the largest methane emission rates will occur around the middle of this century when simulated thaw lake extent is at its maximum and when abrupt thaw under these lakes is taken into account (Schneider von Deimling et al., 2015). Furthermore, the simulated methane fluxes can cause up to 40% of total permafrost-affected radiative forcing in this century. Similarly, no global models currently consider the effects of warming on methane emissions from coastal and ocean shelf systems in the Arctic.

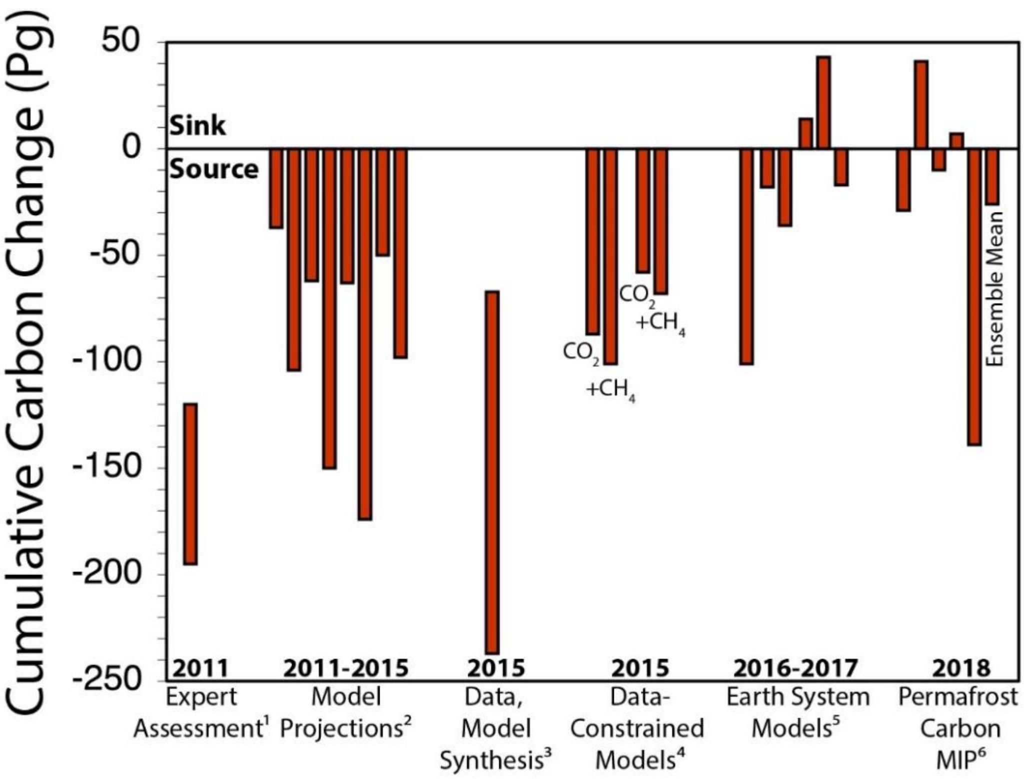


Figure 3.11. Estimates of cumulative net soil carbon pool change for the northern circumpolar permafrost zone by 2100 following the RCP8.5 warming scenario. Cumulative carbon amounts are shown in Petagrams C (1 Pg C=1 billion metric tons), with source (negative values) indicating net carbon movement from soil to the atmosphere and sink (positive values) indicating the reverse. Data are from [1]Schuur et al. 2011 Nature Comment; Schuur et al. (2013); [2]Schaefer et al. (2014)[8 models]; [3](Schuur et al., 2015); [4](Koven et al., 2015); (Schneider von Deimling et al., 2015); [5](MacDougall and Knutti, 2016); (Burke et al., 2017a); [6](McGuire et al., 2018)

3.4.3.1.2 Energy budget

Warming-induced reductions in the duration and extent of Arctic spring snow cover (Section 3.4.1.1) lower albedo because snow-free land reflects much less solar radiation than snow. The corresponding increase in net radiation absorption at the surface constitutes a positive feedback to global climate (Flanner et al., 2011; Qu and Hall, 2014; Thackeray and Fletcher, 2016) (*very high confidence*). The influence of snow albedo on the planetary global energy budget can be quantified using the snow shortwave radiative effect. From 1979 to 2008, changes in snow cover led to an increase in global net solar energy flux at both the surface and top of atmosphere estimated to be 0.22 W m^{-2} (±50%) (*medium confidence*) which weakened the hemispheric top of atmosphere snow shortwave radiative effect (Flanner et al., 2011). Trends in snow shortwave radiative effect during recent decades are weak (Chen et al., 2015; Singh et al., 2015; Chen et al., 2016). A key source of uncertainty in these calculations is the range in observed spring snow cover extent trend estimates (Hori et al., 2017). Terrestrial snow changes also affect the longwave energy budget via altered surface emissivity (Huang et al., 2018).

While projected reductions in snow cover will generally lead to an overall positive climate feedback due to declining albedo, regional variations are also influenced by vegetation (Loranty et al., 2014). There is *medium confidence* in the net effect of potential land cover feedbacks because they may be positive or negative, and will be modulated by many regionally varying factors including: concurrent changes in vegetation distribution (Abe et al., 2017), moisture availability (Myers-Smith et al., 2015; Walker et al., 2015b; Tei et al., 2017), disturbance from fire (Beck et al., 2011), vegetation changes due to permafrost thaw (Helbig et al., 2016a), and associated impacts on latent and ground heat fluxes via canopy shading (Fisher et al., 2016). Since permafrost is a heat sink, decreased permafrost extent, and increased permafrost temperature and active layer thickness (see Section 3.4.1.2) will increase surface temperature and turbulent heat fluxes (Lund et al., 2014). The net effect of these processes remain uncertain.

3.4.3.2 Ecosystems and their Services

3.4.3.2.1 Vegetation

Changes in tundra vegetation can have important ecosystem effects, in particular on hydrology, carbon and nutrient cycling, and surface energy balance, which together impact permafrost (e.g., Myers-Smith and Hik, 2013; Frost and Epstein, 2014; Nauta et al., 2014). Aside from physical impacts, changing vegetation influences the diversity and abundance of herbivores (e.g., Fauchald et al., 2017b; Horstkotte et al., 2017) in the Arctic. The overall trend for tundra vegetation in the 36–year satellite record (1982–2017) shows increasing aboveground biomass (=greening) throughout a majority of the circumpolar Arctic (*high confidence*) (Xu et al., 2013a; Ju and Masek, 2016; Bhatt et al., 2017). The North Slope of Alaska, the Low Arctic (southern tundra subzones) of the Canadian tundra, and east of the Taimyr Peninsula in north-central Siberia, Russia are regions showing the greatest increases. Increasing greenness has been linked with shifts in plant species dominance away from graminoids (grass-like plants) towards shrubs (*high confidence*) (Myers-Smith et al., 2015). Within the overall trend of greening, some tundra show declines in vegetation biomass (=browning) including the Yukon-Kuskokwim Delta of western Alaska, the High Arctic of the Canadian Archipelago, and the northwestern Siberia (Bhatt et al., 2017).

The special variation in greening and browning trends in tundra are also not consistent over time (decadal scale), suggesting interactions between the changing environment and the biological interactions that control these trends. There is *high confidence* that increases in summer, spring, and winter temperatures lead to tundra greening, as well as increases growing season length (e.g., Vickers et al., 2016; Myers-Smith and Hik, 2018) that are in part linked to reductions in Arctic Ocean sea-ice cover (Bhatt et al., 2017; Macias-Fauria et al., 2017). Other factors that stimulate tundra greening include increases in snow water equivalent and soil moisture (Westergaard-Nielsen et al., 2017), increases in active layer depth (via nutrient availability), changes in herbivore activity, and to a lesser degree, human use of the land (e.g., Salmon et al., 2016; Horstkotte et al., 2017; Martin et al., 2017; Yu et al., 2017). Changes in the phenology of tundra vegetation are also documented, and these are linked to changes in snow cover and snowmelt but also interact with other environmental factors like temperature (Oberbauer et al., 2013; Bhatt et al., 2017; Prevéy et al., 2017). Research on tundra browning is more limited but suggests causal mechanisms that include: changes in winter climate—specifically reductions in snow cover due to winter warming events that expose tundra to subsequent freezing and desiccation—insect and pathogen outbreaks, increased herbivore grazing, and ground ice melting and subsidence that increases surface water (Phoenix and Bjerke, 2016; Bjerke et al., 2017) (*medium confidence*).

Similar to tundra, boreal forest vegetation shows trends of both greening and browning over multiple years in different regions across the satellite record (Beck and Goetz, 2011; Ju and Masek, 2016) (*high confidence*). Here, patterns of changing vegetation are a result of direct responses to changes in climate (temperature, precipitation, seasonality) and other driving factors for vegetation (nutrients, disturbance) similar to what has been reported in tundra. Changes in fire disturbance are leading to shifts in landscape distribution of early and late successional ecosystem types, which is also a major factor in satellite trends. Fires that burn deeply into the organic soil layer can alter both physical and biological controls over carbon cycling, including permafrost stability, hydrology, and vegetation. Reduction or loss of the soil organic layer decreases ground insulation (Shur and Jorgenson, 2007; Jorgenson, 2013; Jorgenson et al., 2013; Jiang et al., 2015), warming permafrost soils and exposing old organic matter to microbial decomposition (Schuur et al., 2008). In addition, loss of the soil organic layer exposes mineral soil seedbeds (Johnstone et al., 2009), leading to recruitment of deciduous tree and shrub species that do not establish on organic soil (Kasischke and Johnstone, 2005). This recruitment has been shown to shift post-fire vegetation to alternate successional trajectories (Johnstone et al., 2010). Model projections suggest that Alaskan boreal forest soon may cross a point where recent increases in fire activity have made deciduous stands as abundant as spruce stands on the landscape (Mann et al., 2012). In Arctic larch (*Larix spp.*) forests of northeastern Siberia, increased fire severity may lead to increased tree density in forested areas and forest expansion into tundra as a result of shifting competitive balance between trees and understory or tundra (Alexander et al., 2012).

Fire also appears to be expanding as a novel disturbance into tundra and forest-tundra boundary regions previously protected by cool, moist climate (Jones et al., 2009; Hu et al., 2010; Hu et al., 2015) (*medium confidence*). The annual area burned in Arctic tundra is generally small compared to the forested boreal biome. However, the expansion of fire into tundra that has not experienced large-scale disturbance for centuries causes large reductions in soil carbon stocks (Mack et al., 2011), shifts in vegetation composition and productivity (Bret-Harte et al., 2013), and can lead to widespread permafrost degradation (Jones et al., 2015a). In Alaska – the only region where estimates of burned area exist for both boreal forest and tundra vegetation types – tundra burning averaged approximately 0.3 million ha per year during the last half century (French et al., 2015), accounting for 12% of the average annual area burned throughout the state. Change in the rate of tundra burning projected for this century is highly uncertain as discussed earlier in this chapter (Section 3.4.2.2) (Rupp et al., 2016), but these regions appear to be particularly vulnerable to climatically induced shifts in fire activity. Modelled estimates range from a reduction in activity based on a regional process-model study of Alaska (Rupp et al., 2016) to a fourfold increase across the circumboreal region estimated using a statistical approach (Young et al., 2016).

3.4.3.2.2 Wildlife

Reindeer and caribou (*Rangifer tarandus*), through their numbers and ecological role as a large-bodied herbivore, are a key driver of Arctic ecology. The seasonal migrations that characterize Rangifer link the coastal tundra to the continental boreal forests for some herds, while others live year-round on the tundra. Population estimates and trends exist for most herds, and indicate that Pan-arctic migratory tundra Rangifer have declined from about 5 million in the 1990s to about 2 million in 2017 (Gunn, 2016; Fauchald et al., 2017a) https://carma.caff.is/herds) (*high confidence)*. Numbers have recently increased for two Alaska herds and the Porcupine Caribou herd straddling Yukon and Alaska is at a historic high.

There is *low confidence* in understanding the complex drivers of observed Rangifer decline. Hunting and predation (the latter exacerbated by modification of the landscape for exploration and resource extraction (Dabros et al., 2018); increase in importance as populations decline. Climate strongly influences productivity: extremes in heat, drought, winter icing, and/or deep snow reduce survival (Mallory and Boyce, 2017). Summer heat with drought and/or increased insect parasitism is detrimental to Rangifer survival. Summer warming is changing the composition of tundra plant communities, modifying the relationship between climate, forage, and Rangifer (Albon et al., 2017) (also relevant for other Arctic species such as musk ox; *Ovibos moschatus*; Schmidt et al. (2015)). As polar trophic systems are highly connected (Schmidt et al., 2017), changes will propagate through the ecosystem with effects on other herbivores such as geese and voles, as well as predators such as wolves (Hansen et al., 2013; Klaczek et al., 2016).

Changes in the timing of sea ice formation have direct effects on Rangifer migration and survival. Rangifer in the Canadian Arctic depend on sea ice for inter-island movement and migration to the mainland. For example, sea ice now forms 8–10 days later than it did in the early 1980s between Victoria Island and the mainland, so caribou of the Dolphin and Union herd now cross the strait when the ice is forming which increases risks (Poole et al., 2010). For Eurasian semi-domestic reindeer, late ice formation on waterbodies can impact herding activities (Turunen et al., 2016). Ice formation from rain-on-snow events (Langlois et al., 2017) and dense snow layers from high wind speeds (Dolant et al., 2018) are associated with population changes, including cases of catastrophic mass starvation (Bartsch et al., 2010; Forbes et al., 2016).

In northern Fennoscandia, there are approximately 600,000 semi-domesticated reindeer. Lichen rangelands are key to sustaining reindeer carrying capacity in Fennoscandia and northern Russia. There is variable response of lichen to climate change; enhanced summer precipitation increases lichen biomass, while an increase in winter precipitation lowers it (Kumpula et al., 2014). Fire disturbance reduces the amount of pasture available for domestic reindeer and increases predation on herding lands (Lavrillier and Gabyshev, 2017).

Other non-subsistence terrestrial wildlife can be impacted by observed and projected cryosphere change with some notable effects on predator-prey population cycles, also including temporal and spatial decoupling of trophic interactions. Warmer and shorter winters may have contributed to collapses in the well-known high-amplitude lemming (*Lemmus* spp.; *Dicrostonyx* spp.) population cycles observed in extensive regions of the Arctic (Ims et al., 2008; Kausrud et al., 2008; Ims et al., 2011; Schmidt et al., 2012). These collapses have led to varying degree of concomitant collapses in lemming predator populations (primarily Arctic fox (*Vulpes lagopus*), stoat (*Mustela erminea*), long-tailed jaeger (*Stercorarius longicaudus*) and snowy owl (*Bubo scandiacus*)) depending on how specialized individual predator species are on lemmings as prey, which together may impact the entire ecosystem (Ims and Fuglei, 2005; Gilg et al., 2012; Schmidt et al., 2012). In parallel to population-dependent effects as just described, temporal decoupling of trophic interactions can negatively affect both vegetation and wildlife. For example, warming-induced changes in phenology (timing) of plant flowering and insect emergence is not always synchronous. In Northeast Greenland, the landscape-level flowering season of the dominant plant species was shortened following earlier snowmelt. Earlier snowmelt did not shift the phenology of pollinator insect species, leading to reduced food availability and lower pollinator abundance (Høye et al., 2013; Schmidt et al., 2016; Loboda et al., 2017). In some animals, decreased food at particular stages of development can trigger a cascade of effects in other parts of the animal life cycle that ultimately affects populations levels. For example, the Siberian red knot (*Calidris canutus canutus*), a migratory Arctic shorebird, has been shown to be negatively affected by earlier snowmelt, which reduces food availability for young birds (van Gils et al., 2016). Reduced food may directly affect survival, but has also been shown to result in adult birds with shorter bills. The impact is realized for these birds on their wintering grounds in West Africa, where the shorter bills reduces their ability to reach the high quality and abundant bivalve (*Loripes lucinalis*) buried in the sediments of the intertidal flats. They predominantly rely on the more shallowly buried, but poor-quality, seagrass (*Zostera noltii*) rhizomes, which in turn leads to higher mortality on the wintering grounds. This may be the explanation for declining numbers in the entire flyway population of this species during recent decades. These more subtle but important effects are in addition to longer migration distances, including more stops for rebuilding of body stores, experienced by many migrant bird species following displacement of usable breeding and wintering areas as ecosystem distribution changes across the landscape (Howard et al., 2018). These are just a couple examples of many organismal interactions that demonstrates the far-reaching consequences of changes in polar regions, as well as the difficulty in forecasting the many consequences of change in this region.

3.4.3.2.3 Freshwater

Changes in riparian vegetation (increases of birch, willow, alder replacing other vegetation types) along Arctic river corridors ('shrubification'; Tape et al., 2006; Myers-Smith et al., 2015) enhances inputs of terrestrial nitrogen and carbon from outside aquatic systems into stream networks, stimulating food webs and increasing the productivity of microbial decomposers and invertebrate detritivores (Wrona et al., 2016) (*high confidence*). The role of changing water sources with respect to land ice, snowmelt, and groundwater will influence biological communities (Blaen et al., 2014a). In Arctic snow melt dominated streams, the size of the winter snowpack can influence benthic communities and cause significant inter-stream differences in the same year and intra-stream differences from year to year (Docherty et al., 2017). Glacier-fed rivers are presently experiencing sustained (though finite) periods of increased discharge (Liljedahl et al., 2016) leading to more favorable habitats for some invertebrate and fish species (Vincent et al., 2011). Projected increases in baseflow resulting from permafrost thaw and consequent reduced runoff to infiltration ratios, are *likely* to have a similar effect in sustaining seasonal flow and regulating stream temperatures (Walvoord and Kurylyk, 2016). If conditions become less harsh in streams of the Arctic, potentially more species could be supported, but dispersal constraints related to biogeography limit potential colonization (Hotaling et al., 2017).

Changes in permafrost conditions influence water quality (*high confidence*). Thaw slumps, active layer detachments, and peat plateau collapse result in increased surface water connectivity (Connon et al., 2014) and enhanced sediment and solute flux (Kokelj et al., 2013). The transfer of nutrients from land to water (driven by active layer thickening and thermokarst processes; Abbott et al. (2015); Vonk et al. (2015)) is leading to heightened autotrophic productivity in freshwater ecosystems (Wrona et al., 2016). There is *low confidence* on the influence of permafrost changes on dissolved organic carbon. Permafrost thawing and increased depth to permafrost could enhance transmission of dissolved organic carbon to streams (Wickland et al., 2018) facilitating ammonium retention in these systems and resulting in less export to the ocean (Blaen et al., 2014b). Conversely, reduced dissolved organic carbon export could accompany permafrost thaw as (1) water infiltrates deeper and has longer residences times for DOC decomposition (Striegl et al., 2005) and (2) the proportion of groundwater (typically lower in dissolved organic carbon, higher in dissolved inorganic carbon than runoff) to total streamflow increases (Walvoord and Striegl, 2007). Emerging evidence suggests large stores of mercury in permafrost may be released upon thaw, thereby having effects (largely unknown at this point) on aquatic ecosystems (Schuster et al., 2018).

There is *high confidence* that legacy pollutants like black carbon and persistent organic pollutants (e.g., HCHs, PAHs, PCBs) can be transferred downstream and affect water quality (Hodson, 2014). Lakes can become sinks of these contaminants, while important floodplains can be contaminated (Sharma et al., 2015). An extended growing season for plankton and macrophytes affects water quality and aquatic community structure in lake systems. Shortened duration of snow and ice cover (more light absorption, increased nutrient input) is expected to result in higher primary productivity (Hodgson and Smol, 2008; Vincent et al., 2011). Shifts in the surface water balance are also being observed – permafrost thaw is resulting in drying/draining of lakes and wetlands in some areas, and elsewhere is contributing to the creation of thaw collapse lakes and wetlands (see Section 3.4.1.2.2). The limited amount of human development in the Arctic means local sources of chemical pollution were low; increased human activity is *likely* to lead to enhanced local sources of chemicals of emerging Arctic concern, including siloxanes, parabens, flame retardants, and per- and polyfluoroalkyl substances (AMAP, 2016).

Changes to lake ice phenology (freeze-up, break-up, ice cover duration) and thickness will influence the role that lakes play in regional energy and water budgets (Rouse et al., 2005), while also having implications for biogeochemical cycling and the biological productivity of aquatic systems (*high confidence*). Thinning ice on lakes and streams changes overwintering habitat for aquatic fauna, e.g., by impacting winter water volumes and dissolved oxygen levels (Leppi et al., 2016). Changes in ecological productivity in High Arctic lakes are predominantly controlled by variations in ice-cover duration (Griffiths et al., 2017b). Reductions in ice cover may also encourage greater methane emissions from Arctic lakes (Greene et al., 2014; Tan and Zhuang, 2015).

There is *high confidence* that habitat loss or change due to climate change are serious threats to Arctic fishes. Surface water loss, reduced surface water connectivity among aquatic habitats, and changes to the timing and magnitude of seasonal flows (see Section 3.4.1.2) result in a direct loss of spawning, feeding, or rearing habitats (Poesch et al., 2016). Changes to permafrost landscapes, including the transition from surface water-dominated systems to ground water-dominated systems in some regions (Frey and McClelland, 2008) has reduced freshwater habitats available for fishes and other aquatic biota, including the aquatic invertebrates upon which the fish depend for food. Gullying deepens channels (Rowland et al., 2011; Liljedahl et al., 2016) that otherwise may connect lake habitats occupied by fishes. This can lead to loss of surface water connectivity, limit fish access to key habitats, and lower fish diversity (Haynes et al., 2014; Laske et al., 2016). Small connecting stream channels, which are vulnerable to drying, provide necessary migratory pathways for fishes, allowing them to access spawning and summer rearing grounds (Heim et al., 2016; McFarland et al., 2017).

Changes to the timing, duration, and magnitude of high surface flow events in early and late summer threaten Arctic fish dispersal and migration activities (Heim et al., 2016) (*high confidence*). Timing of important life history events such as spawning can become mismatched with changing stream flows (Lique et al., 2016). Changes to the Arctic growing season (Xu et al., 2013a) increases the risk of drying of surface water habitats and poses a potential mismatch in seasonal availability of food in rearing habitats.

Freshwater systems across the Arctic are relatively shallow, and thus are expected to warm (*high confidence*). This may make some surface waters inhospitably warm for cold water species such as Arctic Grayling (*Thymallus arcticus*) and whitefishes (*Coregonus spp*.), or may increase the risk of *Saprolegnia* fungus that appears to have recently spread rapidly, infecting whitefishes at much higher rates in Arctic Alaska than noted in the past (Sformo et al., 2017). High infection rates may be driven by stress or nutrient enrichment from thawing permafrost, which increases pathogen virulence with fish (Wedekind et al., 2010). Warmer water and longer growing seasons will also affect food abundance because invertebrate life histories and production are temperature and degree-day dependent (Régnière et al., 2012). Increased nutrient export from permafrost loss (Frey et al., 2007), facilitated by warmer temperatures, will *likely* increase food resources for consumers, but how that impacts lower trophic levels within food webs remains speculative. There is regional evidence that migration timing has shifted earlier and egg incubation temperatures have risen for Pink Salmon (*Oncorhynchus gorbuscha*), directly related to warming (Taylor, 2007). While long-term, pan-Arctic data on run timing of fishes are limited, phenological shifts could create mismatches with food availability or habitat suitability in both marine and freshwater environments for anadromous species, and in freshwater environments for freshwater-resident species.

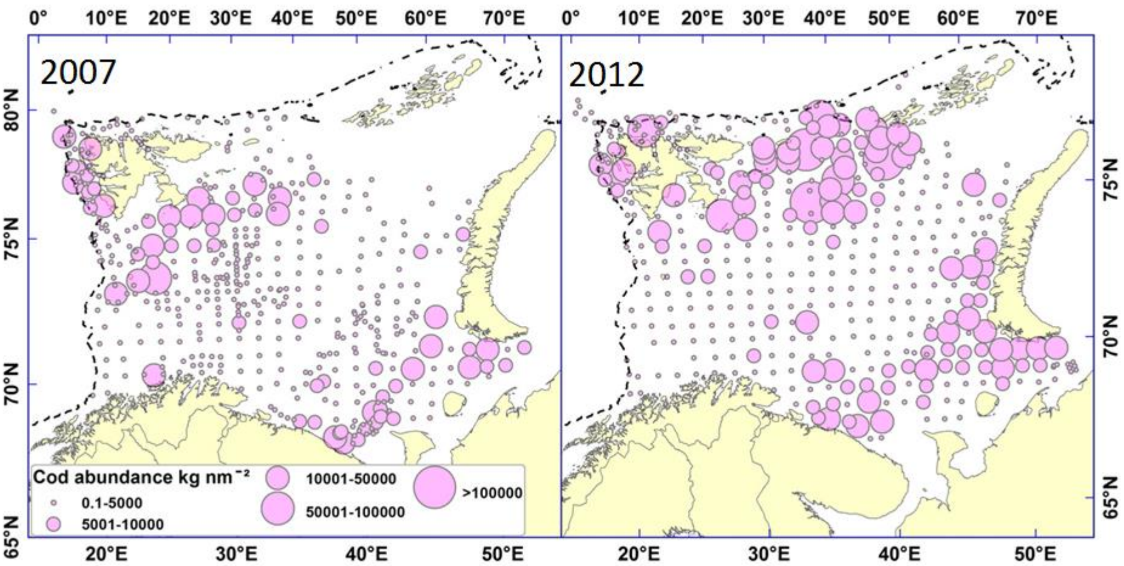
[START BOX 3.3 HERE]

Box 3.3: Impacts and Risks for Polar Biodiversity from Climate Related Range Shifts and Invasive Species

Climate-induced impacts on Arctic and Antarctic marine and terrestrial biodiversity are conveyed through range expansion and human introduction of more temperate species and ecosystems into the polar regions, with higher level of impacts for higher emission scenarios (*high confidence*). The resulting displacements of native species and disruption of ecosystem structure is considered a major threat to biodiversity in both polar regions (CAFF, 2013a; Chown et al., 2017). Substantial differences in geographical settings and physical connectivity to lower latitudes and in the sensitivity of Arctic and Antarctic marine and terrestrial ecosystems, respectively, modify the occurrence and importance of key mechanisms by which such displacements and disruptions occur. Mechanisms include shrinking, degradation and disappearance of polar habitats in favour of more temperate ones, poleward range expansions of lower latitude species encroaching polar ecosystems, and non-native species intentionally or unintentionally brought in by humans becoming invasive, outcompeting native ones.

Species in Arctic marine ecosystems are responding to multiple interacting stresses. Ongoing climate change induced reductions in suitable habitat for Arctic sea ice-affiliated endemic marine mammals is an escalating threat (Section 3.2.3.1), which is complicated by the northward expansion of the summer ranges of a variety of temperate species in the Barents, northern Bering, and Chukchi Seas and increasing pressure from anthropogenic activities. Northward expansions of several whale species have been documented recently in both the Pacific and Atlantic sides of the Arctic (Brower et al., 2017; Storrie et al., 2018). Northward expansion of a range of marine mammals, fishes and seabirds is occuring at the same time as a number of populations of species as different as polar bear (*Ursus maritimus*) and Arctic char (*Salvelinus alpinus*) show range contraction or population declines (Winfield et al., 2010; Bromaghin et al., 2015; Laidre et al., 2018).

Recent studies confirm previous findings that a number of Arctic fish species have changed their spatial distribution patterns substantially over the recent decades (*high confidence*). This may represent the establishment or extirpation (local extinction) of populations in areas that are (now) environmentally suitable or unsuitable. The summer feeding movements of several pelagic and demersal species are impacted by a combination of factors including: changes in suitable habitat availability (Kjesbu et al., 2014; Hu et al., 2015; Eriksen et al., 2017), prey availability, quality, and detection (Varpe et al., 2015; Hunt et al., 2016), the presence of, and consumption by predators (Ingvaldsen and Gjøsæter, 2013), and population density (Kotwicki and Lauth, 2013) (*high confidence*). The sensitivity of some fish species to these multiple stressors, and the relative exposure of species to consequences of climate change, differs by life stage and species (Kjesbu et al., 2014; Barbeaux and Hollowed, 2018). Comparison of ocean conditions in the late 1970s and early 1980s with more recent conditions (2004), in the Barents Sea showed that suitable habitat for two abundant demersal fish stocks (Atlantic cod, *Gadus morhua*, and Northeast Arctic haddock, *Melanogrammus aeglefinus*) extended markedly to the north and east in response to increased sea temperature and retreating sea ice (Ingvaldsen and Gjøsæter, 2013; Landa et al., 2014; Eriksen et al., 2017) (Box 3.3, Figure 1). Similar shifts were also observed in pelagic species, with Barents Sea capelin (*Mallotus villosus*) shifting northwards in recent years (Ingvaldsen and Gjøsæter, 2013) in response to direct and indirect climate effects (Nøttestad et al., 2016). Comparison of the distribution of two dominant subarctic groundfish in the eastern Bering Sea (walleye pollock, *Gadus chalcogrammus*, and Pacific cod, *Gadus macrocephalus*) in a cold (2011) and warm (2017) year revealed both species were distributed farther north in the warm year (Stevenson and Lauth, Submitted).



Box 3.3, Figure 1: Atlantic cod have over the recent years expanded their habitat to the northernmost edge of the Barents Sea. Distribution of cod catches (kilograms per square nautical mile) from bottom trawls during the 2007 (left panel) and 2012 autumn ecosystem surveys. The dashed line indicates the 500 m depth contour. Modified from Kjesbu et al. (2014).

Range expansions have also been observed in the summer feeding distribution of Northeast Atlantic stock of Atlantic mackerel (*Scomber scombrus*), with shifts mainly north westwards into Icelandic and Greenland waters (Jansen et al., 2016; Nøttestad et al., 2016). This range expansion is interpreted to be a result of the continued warming of the ocean in the region (Berge et al., 2015) (*high confidence*). Under RCP4.5 and RCP8.5 further range expansions of Atlantic mackerel are projected in Greenland waters (Jansen et al., 2016). However, these range shifts may be governed by concurrent shifts in mackerel predators such as bluefin tuna (*Thunnus thynnus*) (MacKenzie et al., 2014). Evidence of climate driven spatial shifts in spawn timing or locations is limited to Barents Sea cod (Sundby and Nakken, 2008) and potential shifts are expected to be gradual, responding to factors conducive to survival of young (Höffle et al., 2014; Kvile et al., 2017).

In Arctic marine systems, physical barriers to range expansions into the high Arctic interior shelf systems and the outflow systems of Eurasia and the Canadian Archipelago will continue to govern expansions (*medium confidence*). The limited available information on marine fish from other Arctic Ocean shelf regions reveals a latitudinal cline in the abundance of commercially harvestable fish species (Stevenson and Lauth, 2012). Evidence of latitudinal partitioning between the four dominant mid-water species (Polar cod, *Boreogadus saida*, saffron cod, *Eleginus gracilis*, capelin, and Pacific herring, *Clupea pallasii*) was also observed, with Polar cod being most abundant to the north (Bender et al., 2016). These latitudinal gradients suggest that range expansions of fish species will continue to be governed by a combination of physical factors governing overwintering success and the availability, quality and quantity of prey.

In Antarctic marine systems, alien species introductions are expected to increase (Beaugrand et al., 2015; Fraser et al., 2018), and there is some evidence of range shifts for penguins species on the West Antarctic Peninsula (see Section 3.2.3.2.4). However the ACC and its associated fronts and thermal gradients are expected to persist as a biogeographic barrier for pelagic taxa (Cross-Chapter Box 5) and current evidence of invasions by shell-crushing crabs onto the Antarctic continental shelf remains equivocal (Griffiths et al., 2013; Aronson et al., 2015) (*very low confidence*). Furthermore, as is described in Section 3.2.3.2.2 for Southern Ocean zooplankton, marine assemblages do not necessarily show evidence of changes even in the face of ocean warming.

On Arctic land, northwards range expansions have been recorded in species from all major taxon groups both based on scientific studies and local people's recordings (CAFF, 2013a; AMAP, 2017a; AMAP, 2017b; AMAP, 2018). The most recent examples of vertebrates expanding northwards are a whole range of mammals in Yakutia, Russia (Safronov, 2016), moose (*Alces alces*) in the Arctic regions of both northern continents (Tape et al., 2016) and North American beaver (*Castor canadensis*) in Alaska (Tape et al., 2018). In parallel with these expansions, pathogens and pests are spreading north too (CAFF, 2013a; Taylor et al., 2015; Forde et al., 2016; Burke et al., 2017b; Kafle et al., 2018). An extensive and marked change is so-called tundra greening with subarctic trees and woody shrubs becoming more dominant in Arctic flora and fauna as conditions become more favorable for them (Xu et al., 2013a; CAFF, 2013a; Myers-Smith et al., 2015; Ju and Masek, 2016; Bhatt et al., 2017; Miller et al., 2017; Myers-Smith and Hik, 2018).

Expansion of subarctic species and ecosystems into the Arctic and displacing native species is considered one of the major threats to Arctic species and ecosystems following climate change, since unique Arctic species and ecosystems may be less competitive than encroaching subarctic species favoured by changing climatic conditions (CAFF, 2013a). Similar displacements may take place within the Arctic when Low Arctic species expand into the Mid Arctic, and Mid Arctic species expand into the High Arctic. Here, the most vulnerable species and ecosystems may be in the species poor but unique northernmost sub-zone of the Arctic, because this zone cannot expand northwards itself as southern species and ecosystems are moving in (CAVM Team, 2003; Walker et al., 2016; AMAP, 2018). This 'Arctic squeeze' is a combined effect of the fact that the area of the globe increasingly shrinks when moving north from the Equator and that there is nowhere to go for terrestrial biota when the northern coasts are met. Only on a number of islands in the Arctic Ocean together with high mountain areas is there some room for expansion for the High Arctic sub-zone, but both options offer very little space. The expected overall result of these shifts and limits will be a loss of global biodiversity (CAFF, 2013a; CAFF, 2013b; AMAP, 2018) *(medium confidence)*.

At the southern limit of the Arctic, thermal hotspots may support high biological productivity, but not necessarily high biodiversity (Walker et al., 2015a) and may even act as advanced bridgeheads for expansion of subarctic species into the true Arctic *(medium confidence)*. At the other end of the Arctic zonal range, a temperature increase of only 1°C–2°C in the northernmost subzone may allow the establishment of woody dwarf shrubs, sedges and other species that may radically change its appearance and ecological functions (Walker et al., 2015a) *(medium confidence)*.

On top of climate effects, a multitude of more direct anthropogenic stressors interact to facilitate northward expansion, by provision of additional food resources (e.g., rubbish dumps and roadkill available to predators) or on the contrary to cause declining populations, e.g., due to harvest of migratory birds and mammals inside as well as outside of the Arctic (CAFF, 2013a).

Deeply embedded in the range expansions is the threat from alien species brought in by man to become invasive and outcompete native species. Relatively few invasive alien species are presently well established in the Arctic, but many are thriving on the 'doorstep' in the subarctic and may expand as a result of climate amelioration (CAFF, 2013a; CAFF, 2013b). Examples of this are American mink (*Neovison vison*) and Nootka lupin (*Lupinus nootkatensis*) in western Eurasian and Greenland Arctic, that are already causing severe problems to native fauna and flora, e.g., in Iceland (CAFF and PAME, 2017).

Alien species are a major driver of terrestrial biodiversity change in Antarctica (Chown et al., 2012). The Protocol on Environmental Protection to the Antarctic Treaty prohibits the introduction of non-native species to Antarctica as do the management authorities of sub-Antarctic islands (see De Villiers et al., 2006). Despite this, alien species and their propagules continue to be introduced to the Antarctic continent and sub-Antarctic islands via anthropogenic and natural means (Houghton et al., 2016). To date, 24 insect and plant species have established somewhere in the region (Duffy et al., 2017). Species distribution models for terrestrial invasive species indicate that climate does not currently constitute a barrier for establishment of invasive species on all sub-Antarctic islands, and that the Antarctic Peninsula region will be the most vulnerable location on the Antarctic continent to invasive species establishment under future environmental conditions (under RCP8.5; Duffy et al. (2017)). Thus, for continental Antarctica, existing climatic barriers to alien species establishment will weaken as warming continues across the region (*medium confidence*). An increase in the ice-free area linked to glacier retreat in Antarctica is expected to increase the area available for new terrestrial ecosystems (Lee et al., 2017a), and, along with growing tourist and science visitor numbers, is expected to result in an increase in the establishment probability of terrestrial alien species (Hughes et al., 2015) (*medium confidence*).

[END BOX 3.3 HERE]

3.4.3.3 Impacts on Social-Ecological Systems

The Arctic is home to over four million people, with large regional variation in population distribution and demographics (Heleniak, 2014). 'Connection with nature' is a defining feature of Arctic identity (Schweitzer et al., 2014) because the lands, waters, and ice that surround communities evoke a sense of home, freedom, and belonging and are crucial for culture, life, and survival (Cunsolo Willox et al., 2012; Durkalec et al., 2015). Climate-driven environmental changes are affecting local ecosystems and influencing travel, hunting, fishing, and gathering practises. This has implications for people's livelihoods, cultural practices, economies, and self-determination.

3.4.3.3.1 Subsistence harvesting and food security

Impacts of climate change on food and water security in the Arctic can be severe in regions where infrastructure (including ice roads), travel, and subsistence practices are reliant on elements of the cryosphere such as snow cover, permafrost, and freshwater or sea ice (Cochran et al., 2013; Inuit Circumpolar Council, 2015).

Food security

There is *high confidence* in indicators that food insecurity risks are on the rise for Arctic peoples. Food systems in northern communities are intertwined with northern ecosystems because of traditional and subsistence hunting, fishing, and gathering activities. Environmental changes to animal habitat, population sizes, and movement mean that important food species may no longer be found within accessible ranges or familiar areas (Parlee and Furgal, 2012; Rautio et al., 2014; Inuit Circumpolar Council, 2015; Lavrillier et al., 2016) (Section 3.4.3.2.2). This negatively impacts the accessibility of culturally important local food sources (Lavrillier, 2013; Rosol et al., 2016) that make important contributions to a nutritious diet (Donaldson et al., 2010; Hansen et al., 2013; Dudley et al., 2015). Rain on snow events are a particular challenge for caribou and reindeer to access forage (see 3.4.3.2.1) (Hansen et al., 2014; Overland et al., 2017a; Overland et al., 2017b) with impacts on animal health, mortality, and meat quality in commercial reindeer herding operations (Hansen et al., 2014). Longer open water seasons and poorer ice conditions on lakes impact fishing options (Laidler, 2012) and waterfowl hunting (Goldhar et al., 2014). Permafrost warming and increases in active layer thickness (Section 3.4.1.3) reduce the reliability of permafrost for natural refrigeration. In some cases these changes have reduced access to and consumption of locally resourced food, and can result in increased incidence of illness (Laidler, 2012; Cochran et al., 2013; Cozzetto et al., 2013; Rautio et al., 2014; Beaumier et al., 2015). These consequences of climate change are intertwined with processes of globalization, whereby complex social, economic and cultural factors are contributing to a dietary transformation from locally resourced foods to imported market foods across the Arctic (Harder and Wenzel, 2012; Parlee and Furgal, 2012; Nymand and Fondahl, 2014; Beaumier et al., 2015). Food is strongly tied to culture, identity, values, and ways of life (Donaldson et al., 2010; Cunsolo Willox et al., 2015; Inuit Circumpolar Council, 2015); thus impacts to food security go beyond access to food and physical health.

There is *high confidence* that changes to travel conditions impact food security through access to hunting grounds. Shorter snow cover (Section 3.4.1.1), and changes to snow conditions (such as density), and earlier ice break-up (Section 3.4.1.2) make overland travel more difficult and dangerous (Ford and Pearce, 2012; Laidler, 2012; Cunsolo Willox et al., 2013; Overland et al., 2017b). Changes in dominant wind direction and speed reduce the reliability of traditional navigational indicators such as snow drifts, increasing safety concerns (Ford and Pearce, 2012; Laidler, 2012; Ford et al., 2013; Clark et al., 2016b). Permafrost warming, increased active layer thickness (Section 3.4.1.3), fire disturbance, and changes to water levels (Section 3.4.1.2) impact overland navigability in summer (Goldhar et al., 2014; Brinkman et al., 2016). Of particular concern for coastal communities is landfast sea ice (Section 3.3.1.1.5), which creates an extension of the land in winter that facilitates travel (Inuit Circumpolar Council Canada, 2014). In particular, the floe edge position, timing and dynamics of freeze-up and break-up, sea ice stability through the winter, and length of the summer open water season are important indicators of changing ice conditions and safe travel (Gearheard et al., 2013; Eicken et al., 2014; Baztan et al., 2017). Warming water temperature, altered salinity profiles, snow properties, changing currents and winds all have consequences for the use of sea ice as a travel or hunting platform (Hansen et al., 2013; Eicken et al., 2014; Clark et al., 2016a).

There is *high confidence* that both risks and opportunities arise for coastal communities with changing sea ice and open water conditions. More leads (areas of open water), especially in the spring, can mean more hunting opportunities such as whaling off the coast of Alaska (Hansen et al., 2013; Eicken et al., 2014). In Nunavut, a floe edge closer to shore improves access to marine mammals such as seals or narwhal (Ford et al., 2013). However, these conditions also hamper access to coastal or inland hunting grounds (Hansen et al., 2013; Durkalec et al., 2015), have increased potential for break-off events at the floe edge (Ford et al., 2013), or can result in decreased presence (or total absence) of ice-associated marine mammals with an absence of summer sea ice (Eicken et al., 2014).

Water security

Many northern communities rely on ponds, streams, and lakes for drinking water (Cochran et al., 2013; Goldhar et al., 2013; Nymand and Fondahl, 2014; Cunsolo Willox et al., 2015; Daley et al., 2015; Dudley et al., 2015; Overland et al., 2017b), so there is *high confidence* that projected changes in hydrology will impact water supply (Section 3.4.2.2). Surface water is vulnerable to thermokarst disturbance and drainage, as well as bacterial contamination, both of which are impacted by warming ground and water temperatures (Cozzetto et al., 2013; Goldhar et al., 2013; Dudley et al., 2015; Overland et al., 2017b; Wright et al., 2017). Icebergs or old multi-year ice are important sources of drinking water for some coastal communities, so changing accessibility affects local water security. Some small remote communities have limited capacity to respond quickly to water supply threats, which amplifies vulnerabilities of water security (Daley et al., 2015).

3.4.3.3.2 Communities

Culture and knowledge

Spending time on the land is culturally important for Indigenous communities (Eicken et al., 2014; Durkalec et al., 2015; Inuit Circumpolar Council, 2015). There is *very high confidence* that climate change impacts daily life because land-based activities and community events are closely tied to seasonal cycles connected to ice freeze-up and break-up (rivers/lakes/sea ice), snow onset/melt, vegetation phenology, and related wildlife/fish/bird behaviour (Inuit Circumpolar Council, 2015). Inter-generational knowledge transmission of associated values and skills is also influenced by climate change (Ford and Pearce, 2012; Eicken et al., 2014; Cunsolo Willox et al., 2015; Inuit Circumpolar Council, 2015). Where changes are happening rapidly or unpredictably, younger generations in the Canadian Arctic do not have the same level of experience or confidence with traditional indicators (Ford, 2012; Parlee and Furgal, 2012; Cunsolo Willox et al., 2015). This threatens confidence in Indigenous Knowledge holders (Ford and Pearce, 2012; Parlee and Furgal, 2012; Cunsolo Willox et al., 2015; Golovnev, 2017).

Economics

The northern mixed economy is characterized by a combination of subsistence activities, and employment and cash income. The social economy related to sharing, kinship, and the framing of household economic conditions has received limited research attention (Ford and Pearce, 2012; Harder and Wenzel, 2012; Fall, 2016) It is difficult to assess the impact of climate change on local subsistence activities and economic opportunities (e.g., fishing, resource extraction, tourism and transportation) because of high variability between communities (Cunsolo Willox et al., 2012; Ford and Pearce, 2012; Harder and Wenzel, 2012; Cochran et al., 2013; Fall, 2016; Ford et al., 2016; Clark et al., 2016b). Longer ice-free travel windows in Arctic seas could lower the costs of access and development of northern resources (delivering supplies and shipping resources to markets) and thus, may contribute to increased opportunities for marine shipping, commercial fisheries, tourism, and resource development (Ford et al., 2012; Huskey et al., 2014; Overland et al., 2017b). This has important implications for economic development, particularly in relation to local employment opportunities but raises also concerns of detrimental impacts on animals, habitat, and subsistence activities (Cochran et al., 2013; Inuit Circumpolar Council, 2015). There are many marine transport risks associated with unpredictable sea ice conditions, and development costs could remain high due to increased flooding, coastal erosion, and impacts on infrastructure (Huskey et al., 2014).

3.4.3.3.3 Health and wellbeing

For many polar residents, especially Indigenous peoples, the physical environment underpins social determinants of well-being, including physical and mental health. Changes to the environment impact most dimensions of health and well-being (Parlee and Furgal, 2012; Driscoll et al., 2013). Climate change consequences in polar regions (Section 3.3.1.1; 3.4.1.2) have impacted key transportation routes (Gearheard et al., 2006; Laidler, 2006; Ford et al., 2013; Clark et al., 2016a) and pose increased risk of injury and death during travel (Driscoll et al., 2013; Durkalec et al., 2014; Durkalec et al., 2015; Driscoll et al., 2016; Clark and Ford, 2017).

Foodborne disease is an emerging concern in the Arctic because warmer waters, loss of sea ice (Section 3.3.1.1) and resultant changes in contaminant pathways can lead to bioaccumulation and biomagnification of contaminants in key food species. While many hypothesized foodborne diseases are not well studied (Parkinson and Berner, 2009), foodborne gastroenteritis is associated with shellfish harvested from warming waters (McLaughlin et al., 2005; Young et al., 2015).

Climate change increases the risk of waterborne disease in the Arctic via warming water temperatures and changes to surface hydrology (Section 3.4.1.2) (Parkinson and Berner, 2009; Brubaker et al., 2011; Dudley et al., 2015). After periods of rapid snowmelt, bacteria can increase in untreated drinking water, with associated increases in acute gastrointestinal illness (Harper et al., 2011). Consumption of untreated drinking water may increase duration and frequency of exposure to local environmental contaminants or potential waterborne diseases (Goldhar et al., 2014; Daley et al., 2015). The potential for infectious gastrointestinal disease is not well understood, and there may be greater concerns in relation to storage containers of raw water than the source water itself (Goldhar et al., 2014; Wright et al., 2017).

Climate change has negatively affected place attachment via hunting, fishing, trapping, and traveling disruptions, which have important mental health impacts (Cunsolo Willox et al., 2012; Durkalec et al., 2015; Cunsolo and Ellis, 2018). The pathways through which climate change impacts mental wellness in the Arctic varies by gender (Bunce and Ford, 2015; Harper et al., 2015; Bunce et al., 2016) and age (Petrasek-MacDonald et al., 2013; Ostapchuk et al., 2015). Emotional impacts of climate-related changes in the environment were significantly higher for women compared to men, linked to concern for family members (Harper et al., 2015). However, men are also vulnerable due to gendered roles in subsistence and cultural activities (Bunce and Ford, 2015). In coastal areas, sea ice means freedom for travel, hunting, and fishing, so changes in sea ice affect the experience of and connection with place. In turn, this influences individual and collective mental/emotional, spiritual, social, and cultural health according to relationships between sea ice use, culture, knowledge, and autonomy (Gearheard et al., 2013; Durkalec et al., 2015; Inuit Circumpolar Council, 2015).

3.4.3.3.4 Infrastructure

Permafrost is undergoing rapid change (Section 3.4.1.3), creating challenges for planners, decision makers, and engineers (AMAP, 2017). The observed changes in ground thermal regime (Romanovsky et al., 2010; Romanovsky et al., 2017a; Romanovsky et al., 2017b) threaten the structural stability and functional capacities of infrastructure (defined here as facilities with permanent foundations on ice-free land), in particular in ice-rich frozen ground. Extensive summaries of construction damages along with adaptation and mitigation strategies are available (Instanes et al., 2005; Callaghan et al., 2011; Larsen et al., 2014; Doré et al., 2016; Pendakur, 2017; Vincent et al., 2017; Shiklomanov et al., 2017a; Shiklomanov et al., 2017b). Although engineering solutions can address both human-induced and naturally caused infrastructure challenges, their economic cost may be prohibitive at regional scales (Doré et al., 2016). Thus, broad-scale knowledge on hazardous environments and magnitude of potential infrastructure risks are of importance for planners and policy-makers in the coming decades (AMAP, 2017).

Under RCP4.5, it is *likely* that approximately 70% of circumpolar infrastructure (residential, transportation and industrial facilities), including over 1200 settlements (~40 with population more than 5000) are located in areas where permafrost is projected to thaw by 2050 (Hjort et al., submitted). Regions associated with the highest hazard are in the thaw-unstable zone characterized by relatively high ground-ice content and thick deposits of frost-susceptible sediments (Shiklomanov et al., 2017b). By 2050, these high-hazard environments contain one-third of existing pan-Arctic infrastructure (Hjort et al., submitted).

Onshore hydrocarbon extraction and transportation in the Russian Arctic are at risk: 45% of the oil and natural gas production fields in the Russian Arctic are located in the highest hazard zone. Critical areas in future decades include the Pechora region, northwestern parts of the Ural Mountains, and northwest and central Siberia (Instanes, 2016; Shiklomanov et al., 2017b; Hjort et al., submitted). Reducing greenhouse gas emissions under a scenario roughly consistent with the Paris Agreement (RCP2.6), could stabilize potential risks to infrastructure after mid-century. In contrast, high emission scenarios (RCP8.5) would result in continued negative climate-change impacts on the built environment and economic activity in the Arctic (Hjort et al., submitted).

For the state of Alaska, cumulative expenses projected for climate-related damage to infrastructure totalled USD5.5 billion between 2015 and 2099 under RCP8.5 (Melvin et al., 2017). The top two causes of damage related costs were projected to be road flooding from increased precipitation, and building damage associated with near-surface permafrost thaw. These costs decreased by 24% for the same time frame under RCP4.5, indicating that reducing greenhouse gas emissions globally could lessen damages. Adaptation measures reduced damage-related costs by over 50% in both emission scenarios.

Winter roads (snow covered ground and frozen lakes) influence the reliability and costs of transportation to supply and connect remote northern communities and industrial development sites (Parlee and Furgal, 2012; Huskey et al., 2014; Overland et al., 2017b). For travel to and between northern communities, changing lake and river levels and the period of safe ice cover all affect the duration of use of overland travel routes and inland waterways, with associated implications for increased travel risks, time, and costs (Laidler, 2012; Ford et al., 2013; Goldhar et al., 2014). Although ice growth is accelerated for some winter roads by removing overlying snow and flooding with lake water, there have been recent instances of severely curtailed ice road shipping seasons due to unusually warm conditions in the early winter, which prevented this intervention (Sturm et al., 2017). While the impact of human effort on the seasonal development and maintenance of ice roads is difficult to quantify, reduction in the operational time window due to winter warming is projected (Mullan et al., 2017).

3.5 Responding to Climate Change in Polar Systems

This section assesses past and possible future human responses to climate change in polar social-ecological systems. Special attention is given to systems of governance and how they are providing for climate change mitigation, adaptation, and transformation. Pathways for building resilience to climate change are explored through a review of promising strategies currently being employed in Polar Regions.

3.5.1 The Polar Context for Human Responses to Climate Change

Human responses to climate change in polar regions (like other regions) are part of social-ecological processes undertaken concurrently at and across multiple levels (Ford et al., 2014b; Palmer and Smith, 2014; Adger et al., 2018). In the Antarctic, responses to climate change primarily involve polar scientists, logistics support staff, tour operators, fishers, non-governmental organisation (NGO), and national governments interacting at the international arena. In the Arctic, responses are undertaken with a more diverse set of actors at and across local to international decision-making arenas. In both regions, human perception, values, history, and choice all affect human responses, with social, cultural, economic, political and legal systems interacting to shape outcomes. Concurrently, the drivers of climate change interact with other forces for change, such as globalization, land- and sea-use change, and economic change, necessitating an assessment of both cumulative effects and context-specific pathways for achieving positive resilience (Nymand and Fondahl, 2014; ARR, 2016). In both northern and southern high latitudes, extreme climatic conditions and remoteness from densely populated regions constrain human choice. These constraints follow from restricted human mobility, limited biological productivity during warmer seasons, a paucity of baseline data, difficulties and high cost of travel, and a complex geo-political environment. In the Arctic, differing interests and cultural orientations, including those of Indigenous people who view the Arctic as ancestral homelands, non-residents of the Arctic residing in urban environments, resource extraction corporations, nature-based NGOs, tourists, and others also challenge collective responses to climate change (Shadian, 2014; Shadian, 2017). Adding to this dynamic are those people who have never been to the polar regions, but who have concerns for the regions' future. This complexity limits our capacity to explain past responses and predict future responses with certainty.

Human occupancy in Antarctica is relatively recent. In short, Antarctica is no one's homeland, and instead, represents a commons under the stewardship of many countries. The northern latitudes, on the other hand, have for millennia been the homelands of Indigenous peoples. Approximately 4 million people reside in the Arctic, and differ widely by region, ranging from 94% of Iceland's population living urban and 68% Nunavut, Canada's population living in rural areas. And while there is a general movement to greater urbanization in the Arctic populations, that trend is not true for all regions of the North (Heleniak, 2014). And while 'climigration' (migration because of the impacts of climate change) has been discussed (AHDR, 2015), regional empirical studies provide no evidence that it has occurred (Hamilton et al., 2016b).

About 10% of Arctic residents are counted as Indigenous, although determinations of what constitutes 'Indigenous' are disputed (AHDR, 2015; ARR, 2016). For example, 85% of Nunavut, Canada are Indigenous, and about 15% of Alaskans are 'Native' (Fondahl et al., 2015). Ethnicity and cultural orientation do in climate change responses (Adger et al., 2012), as do histories of colonization and the level of regional political autonomy from southern-based nation states policies (Keil and Knecht, 2017). Human responses to climate change in Antarctic, on the other hand, are largely shaped by international agreements and the informal cooperation of the region's stakeholders (see section 3.5.5.2).

Rural residents, and Indigenous Arctic communities in particular continue to be sustained with mixed cash-subsistence economies that are highly dependent on hunting, herding, fish, herding, and gathering (Nymand and Fondahl, 2014). This high dependence on and long relationship with living resources of the ocean and land make Indigenous peoples especially sensitive to climate change in ways that inform understanding of ecological impacts (Huntington et al., 2018), human adaptation (Ford et al., 2015; Pearce et al., 2015) and resource governance (Danielsen et al., 2014; Forbes et al., 2015). And for Indigenous Peoples, human responses to climate change are considered a matter of cultural survival (Greaves, 2016) (see Cross-Chapter Box 3 in Chapter 1). Indigenous people, however, are neither homogenous in their perspectives nor apart from other sectoral activity areas. While in some cases Indigenous People are negatively impacted by sectoral activities such as mining and oil and gas development (Nymand and Fondahl, 2014), in other cases they benefit financially (Shadian, 2014), setting up dilemmas and potential internal conflicts (Huskey, 2018; Southcott and Natcher, 2018) *(high confidence).*

Polar regions are also unique with respect to the novelty of their systems of governance. The Antarctic Treaty, Indigenous land claims and self-governance agreements, the role of Sami Council in Fennoscandia, Russian Association of Indigenous Peoples of the North in Russia, Inuit Circumpolar Council, nature-based NGOs, resource co-management arrangements, and the Arctic Council are a few of the organizational and institutional innovations in polar regions that provide opportunities for responding to climate change *(medium confidence).* Governance is addressed in section 3.5.5.

3.5.2 Assessing Human Responses in Polar Regions: Adaption and Resilience

Recent literature assessing polar climate change has shifted from the study of risk and vulnerability to also examine the resilience of social-ecological systems (see Cross-Chapter Boxes 1 and 2 in Chapter 1). A focus on resilience forces an examination of the dynamic nature of coupled social-ecological interactions, potential SES regime shifts and their respective thresholds of change, and the role (and limitations) of human agency in mitigation, adaptation and transformation (ARIR, 2013; ARR, 2016; AMAP, 2017a). Moreover, considering social-ecological resilience highlights that adaptation is most successful when it addresses the immediate risks and vulnerabilities while concurrently building resilience of the system for possible future conditions (AMAP, 2017; AMAP, 2017a; AMAP, 2018). The resilience frame poses a number of questions such as, how might climate induced regime shifts in polar regions affect the supply of ecosystem services, and in turn, affect human livelihoods and well-being? Are some individuals and groups of polar regions more able to adapt and or transform to climate change than others? What conditions contribute to or impede adaption and transformation? What resources are critical for realizing resilient climate pathways for the future? This assessment considers use of strategies for achieving resilient social-ecological systems (see Cross-Chapter Box 1 in Chapter 1). They include i) maintaining system diversity and redundancy; ii) using a complex systems approach to understand phenomena and problems, iii) encouraging social learning and experimentation; iv) broadening participation in decision making; v) managing slow variables and feedbacks, vi) enhancing polycentric systems of governance and vii) managing connectivity.

We also consider what assets and tools have contributed to adaptive capacity or are lacking. Types of assets and tools include geography (e.g., remoteness from required resources); ecosystem (e.g., diversity of living resources); physical infrastructure (e.g., buildings, roads, communication systems); human capital (e.g., skill of leaders to navigate legal processes); finances (e.g., sources of income for funds to support adaptation); social and cultural capital (e.g., access to social networks to access information and level of trust among group members to act collectively); and institutions (e.g., formal and informal rules, such as property rights, formal agreements, and treaties) (Hovelsrud and Smit, 2010; Kofinas et al., 2013; Kofinas et al., 2016; Berman, 2017). Use of assets are context specific; while one set of assets and tools may be especially critical in one case, e.g., reindeer herders having use of alternative pasturelands to move deer in response responding to rain-on-snow events (Bartsch et al., 2010; Forbes et al., 2016), a different set may be needed by another group, e.g., commercial fishers drawing on a highly responsive resource management system to address changes in the distribution and abundance of fish stocks. As well, having access to assets is no guarantee of their use; awareness of conditions and motivation to act matter (van der Linden et al., 2015) (*medium confidence*).

3.5.3 Human Responses

Table 3.7 summarizes the consequences, interacting drivers, responses, and assets of climate change responses by select social-ecological systems (i.e., sectors) of Arctic and Antarctic regions. Also noted are anticipated future conditions and level of certainty and other drivers of change that may interact with climate and affect outcomes. Implications to world demands on natural resources, innovation and development of technologies, population trends and economic growth are *likely* to affect all systems, as is the global significance of the Paris Climate Agreement (AMAP, 2017b). In several cases, drivers of change interacting with climate change are regionally specific and not easily captured. In many cases there is limited information on human responses to climate change in the Russian Arctic.

3.5.3.1 Fisheries

Responses addressing changes in the abundance and distribution of fish resources differ by region. In some polar regions strategies of adaptive governance, biodiversity conservation, scenario planning, and the precautionary approach are already in use. Further development of coordinated monitoring programs, data sharing, social learning and decision-support tools that alert managers to climate change impacts on species and ecosystems would allow for appropriate and timely responses including changes in overall fishing capacity, individual stock quotas, shifts between different target species, opening/closure of different geographic areas and balance between different fishing fleets. This will contribute to the resilience and conservation of these natural-social systems (*medium confidence*).

Five Arctic States, known as 'Arctic 5' (Canada, Denmark, Norway, Russia and the United States) have sovereign rights for exploring and exploiting resources within their 200 nm EEZs in the High Arctic and manage their resources within their own regulatory measures. A review of future harvest of living resources in the European Arctic by Haug et al. (2017) points towards high probability of increased northern movement of several commercial fish species (Section 3.3.3.1 and Box 3.3), but only to the shelf slope for the demersal species. This suggests increased northern fishing activity, but within the EEZs and present management regimes (Haug et al., 2017) (*medium confidence*).

In Norway's EEZs a new Marine Resources Act entered into force in 2009. This act applies to all wild living marine resources and states that its purpose is to ensure sustainable and economically profitable management of the resources. Conservation of biodiversity is described as an integral part of its sustainable fisheries management and it is mandatory to apply 'an ecosystem approach, taking into account habitats and biodiversity' (Gullestad et al., 2017). A scenario based approach to identify management strategies that are effective under changing climate conditions is also being explored for the Barents Sea (Planque et al., Submitted). In addition to national management, the Joint Norwegian-Russian Fisheries Commission provides cooperative management of the most important fish stocks in the Barents and Norwegian Seas. The stipulation of the total quota for the various joint fish stocks is a key element, as is more long-term precautionary harvesting strategies, better allowing for responses to climate change (*medium confidence*).

In the U.S. Arctic a proactive approach to fishery management has been introduced that utilizes future ecological scenarios to develop strategies for mitigating the future risks and impacts of climate change (NPFMC, 2018). The fisheries of the southeastern Bering Sea are managed through a complex suite of regulations that include catch shares (Ono et al., 2017), habitat protections, restrictions on forage fish, bycatch constraints (DiCosimo et al., 2015), and community development quotas. This intricate regulatory framework has inherent risks and benefits to fishers and industry by limiting flexibility (Anderson et al., 2017b).To address these challenges, the NPFMC recently adopted a Fishery Ecosystem Plan (FEP), which includes a multi-model climate change action module (Punt et al., 2015; Holsman et al., 2017; Zador et al., 2017; Hermann et al., Submitted).

The US has prohibited commercial fishing in their EEZ of the Chukchi and Beaufort Seas until sufficient information is obtained to sustainably manage the resources (Wilson and Ormseth, 2009).In the Canadian sector of the Beaufort Sea, commercial fisheries are until now only small scale and locally operated. However, climate change with decreasing ice cover together with over-harvesting of fish stocks in other places may increase incentives to exploit the resource. This risk has caused concern among local Inuvialuit subsistence fishers of the western Canadian Arctic. In response, a new proactive ecosystem-based Fisheries Management Framework was developed (Ayles et al., 2016). Also in Western Canada, the commercial fishery for Arctic char (*Salvenius alpinus*) in Cambridge Bay is co-managed by local Inuit organizations and Fisheries and Oceans Canada (DFO, 2014).

The high seas region of the Central Arctic Ocean (CAO) is per definition outside of any nations EEZ. Recent actions of the international community shows that a precautionary approach to considerations of CAO fisheries has been adopted (*high confidence*) and that expansion of commercial fisheries into the CAO will be constrained until sufficient information is obtained to manage the fisheries according to an ecosystem approach to fisheries management (*high confidence*). The Arctic 5 have officially adopted the precautionary approach to fishing in 2015 by signing the Oslo Declaration concerning the prevention of unregulated fishing in the CAO. The declaration established a moratorium to limit potential expansion of CAO commercial fishing until sufficient information, also on climate change impacts, is available to manage it sustainably. This was followed up in 2017, when at a global regulatory level, the Arctic 5 and several other nations agreed to a treaty that imposed a 16 year moratorium on commercial fishing in the CAO and encouraged research cooperation (Conservation and sustainable Use of Marine Biodiversity of Areas beyond National Jurisdictions; BBNJ) (Lui, 2017). Several other agreements have adopted the same approach, including the Central Arctic Ocean Fisheries (CAOF) Agreement.

CCAMLR is responsible for the conservation of marine resources south of the Antarctic Polar Front (CCAMLR, 1982) and has ecosystem-based fisheries management embedded within its Convention (Constable, 2011). This includes the CCAMLR Ecosystem Monitoring Program, which aims to monitor important land-based predators of krill to detect the effects of the krill fishery on the ecosystem. Currently, there is no formal mechanism for choosing which data are needed in a management procedure for krill or how to include such data. However, this information will be important in enabling CCAMLR fisheries management to respond to the effects of climate change on krill and krill predators in the future.

The displacement of fishing effort will impact fishing operations in the CAMLR Convention area under future climate change (*medium confidence*). Such displacement could be attributed to both the poleward shifts in species distribution (Pecl et al., 2017), although McBride et al. (2014) noted that the potential for invasion into the Southern Ocean of large and highly productive pelagic finfish appears low) or management techniques establishing marine protected areas, such as the Ross Sea MPA (Brooks, 2013) (*low confidence*). Fisheries in the Southern Ocean operate over large spatial ranges within which conditions are likely to change differently by region. Yet as those fisheries are relatively mobile, they are potentially able to respond to range shifts in target species, which is in contrast to small-scale/coastal fisheries in other regions (*very low* *confidence*). Fishing operations are also impacted by the navigational hazards caused by unpredictable sea-ice conditions and duration (ATCM, 2017), which can serve to change the spatial distribution of fishing operations and their associated management processes (Jabour, 2017).

3.5.3.2 Transportation

Without well-developed management plans and regulations, recent and future increases in polar transportation (i.e., shipping and air travel) will result in greater risk to humans and ecosystems, such as an increased likelihood of accidents, the introduction of invasive species, oil spills, and waste discharges.

Arctic shipping activity, especially in certain geographic areas (NSR, AB and eventually NWP and maybe TPR) has and is *likely* to continue increasing in the future (Stephenson et al., 2011; Smith and Stephenson, 2013; Stephenson et al., 2013). Industry has responded by investing in development of shipping design for travel in mixed ice environments. These increases are occurring in spite of the limited total savings when comparing shorter travel to increased CO_2 emissions (Lindstad et al., 2016). In anticipation of spills, research in several regions have explored oil spill response viability and new methods of oil spill response for the Arctic environment (Bullock et al., 2017; Dilliplaine, 2017; Holst-Andersen et al., 2017) *(medium confidence).* Statoil has developed and uses risk assessment decision-support tools for environmental management, together with environmental monitoring (Utvik and Jahre-Nilsen, 2016). The tools allow for qualification to assess Arctic oil-spill response capability, ice detection in low visibility and improved management of sea ice and icebergs, and numerical modelling of icing and snow as risk mitigation.

The International Maritime Organization is the body responsible for regulating international Arctic shipping. There are a number of mechanisms standardizing regulation and governance (the International Convention for the Prevention of Pollution from Ships, MARPOL; the International Convention for the Safety of Life at Sea, SOLAS; the International Convention on Standards of Training, Certification and Watchkeeping for Seafarers, STCW), including recent Arctic initiatives, such as joint search and rescue agreements and joint oil pollution response, and the newly implemented Polar Code (IMO, 2017). The agreement was consensus based, hence implemented at the lowest common denominator, including a call to enhance enforcement capabilities and address emerging issues such as heavy fuel oil and black carbon, among other environmental protection provisions regulating heavy fuel oil (HFO) transport and use, black carbon, and ballast water (Anderson, 2012; Sakhuja, 2014; IMO, 2017). And while the Polar Code does address emerging issues, it may be deficient in its capacity to meet future needs.

National-level regulation varies (some stronger than others) and ships with flags of convenience can cause challenges (Chircop, 2009; Anderson, 2012; Sakhuja, 2014; IMO, 2017). National-level responses have included several studies to consider scenarios of change and explore regulatory changes. Continued, and in some areas, greater international cooperation on shipping governance would be helpful for addressing emerging climate change issues (Arctic Council, 2015; ARR, 2016; PEW Charitable Trust, 2016; Chénier et al., 2017; IMO, 2017) *(high confidence).*

The IMO Polar Code came into force in 2017 with the purpose of setting new standards for vessels travelling in polar areas to avoid environmental damage and to improve safety (IMO, 2017). The IMO Polar Code, however, currently excludes fishing vessels and vessels on government service, thereby excluding many shipping activities in the Antarctic region (IMO, 2017) *(high confidence).*

The International Maritime Organization Polar Code of 2017 will set new standards for vessels travelling in polar areas to avoid environmental damage and to improve safety (IMO, 2017). The IMO Polar Code, however, currently excludes fishing vessels and vessels on government service, thereby excluding many shipping activities in the Antarctic region (IMO, 2017) because many ships travelling these waters will continue to pose risks to the environment and to themselves as they are not regulated under the Polar Code *(high confidence)*.

With frequent use of ice runways and increase of air traffic by both National Antarctic Programmes and tourism operators, the unpredictable state of airstrips may alter such transportation and the infrastructure to support them (ATCM, 2017). For example, due to current thawing of ice runways and possible future impacts to accessing the Italian Mario Zuchelli station, Italy has proposed constructing a gravel runway (Italy, 2015).

3.5.3.3 Non-renewable Extractive Industries

Climate change has forced, to a limited extent, non-renewable resource extraction industries and agencies that regulate their activities to respond to changes in sea ice, thawing permafrost, spring run offs, and resultant timing of exploration, construction and use of ice roads, and infrastructure design. In some regions, climate change has offered new development opportunities and prompted development of better forecasting tools to help anticipate future conditions.

Exploitation of natural resources in the Antarctic is prohibited by the Antarctic Treaty. In the Arctic, receding sea ice and glaciers has opened new possibilities for development, such as areas of receding glaciers of eastern Greenland (Smits et al., 2017). As oil and gas exploration got underway in Greenland, its home rule government developed environmental impact assessment protocols to provide for adequate public participation (Forbes et al., 2015). On the North Slope of Alaska, oil and gas development is now undergoing new expansion, while industry concurrently faces increasing challenges of climate change, such as shorter and warmer winters, the main season for oil exploration and production (Lilly, 2017). The method for building of ice roads on the North Slope has been somewhat modified to account for warmer temperatures during construction. Flooding events on the North Slope of Alaska due to unusually high spring melt and run off in 2015 closed the Dalton Highway and North Slope oil field operations for an extended period, resulting in financial losses to companies, and suggesting that the need to rethink the design of culverts and roads (Raynolds et al., 2012; Walker et al., Submitted). There are also knowledge gaps in understanding implications of seismic studies with climate change on the landscape (Dabros et al., 2018) The issue of cumulative effects also raises questions of current practice of environmental impact assessment to evaluate potential cumulative effects (Walker et al., Submitted).

Lilly (2017) reported that optimizing Alaska North Slope transportation networks during winter field operations is critical in managing increasing resource development, and could potentially provide a better framework for environmentally-responsive development. Better understanding of environmental change is also important in ensure continued oil field operations with protection of natural resources. Better forecasting of short-term conditions (i.e., snow, soil temps, spring run offs) could allow management agencies to respond to conditions more proactively, and industry more time to plan winter mobilization, such as construction of ice roads (*low confidence*).

3.5.3.4 Arctic Subsistence Systems

Subsistence users have responded to climate change by adapting their wildfood production and engaging in the climate policy processes at multiple levels of governance. The limitations of many formal institutions suggest that in order to achieve greater resilience of subsistence systems, transformations in governance are needed to provide greater power sharing, including more resources for engaging in regional-to-national policy making.

Responses to climate change fall into several categories. In some cases harvesters are shifting the timing of harvesting and the selection of harvest areas due to changes in seasonality and limited access to traditionally used areas. Changes in the navigability of rivers and more open (i.e., dangerous) seas has resulted in harvesters changing harvesting gear, such as shifting to from propeller to jet-propelled boats or all-terrain-vehicles, and to larger ocean-going vessels for traditional whaling (Brinkman et al., 2014). In many cases, using different gear results in an increase in fuel costs (e.g., jet boats are about 30% less efficient). In Savoonga, Alaska, whalers reported limitations in harvesting larger bowhead because of thin ice conditions that do not allow for safe haul outs. As a result, community residents now anticipate a greater dependence on western Alaska's reindeer as a source of meat (Rosales and Chapman, 2015) in future. Harvesters have also responded with switching of harvested species and in some cases doing without (AMAP, 2018). Evidence also shows that in many cases, adaption has allowed for continued provisioning of wildfoods (BurnSilver et al., 2016; Fauchald et al., 2017b; AMAP, 2017a) (*medium confidence*).

The impacts of climate change have also required adaptation to non-harvesting aspects of wildfood production, such as an increased use of household and community freezers, and in some cases, an abandonment of traditional food drying practices. And several cases there has been an increased emphasis on community self-reliance, such as use of household and community gardens for food production (Loring et al., 2016). In the future agriculture may be more possible with improved conditions at the southern limit of the Arctic, and thus supplement hunting and fishing (AMAP, 2018).

Climate change may in the future bring both new harvestable fish, birds, mammals and berry producing plants to the North, and reduced populations and/or access to currently harvested species (AMAP, 2017a; AMAP, 2017b; AMAP, 2018). Adaptive co-management and integration of local to regional level management with national- to international-level agreements necessitates consideration for sustainable harvest of new resources as well as securing sustainable harvest or even full protection of dwindling or otherwise vulnerable populations. In these cases, adaptive co-management can be an efficient tool to secure agreed population goals, including international cooperation and agreements regarding migratory species shared between more countries (Kocho-Schellenberg and Berkes, 2014) (see Section 3.5.4.9).

Indigenous leaders are responding by engaging more actively in political processes on climate change at multiple levels and through different avenues. At the United Nations Framework Convention on Climate Change (UNFCCC), the discursive space for incorporating perspectives of Indigenous People on climate change adaptation has expanded since 2010, which is reflected in texts and engagement with most activity areas (Ford et al., 2015). Aleut International Association, Arctic Athabaskan Council, Gwich'in Council International, Inuit Circumpolar Council, and Russian Association of Indigenous Peoples of the North, and the Saami Council, which sit as ‘permanent participants’ of the Arctic Council, are involved in many of the AC’s working groups. (Sections 3.5.5.2.2 and 3.5.5.2.3). Greater involvement at the national and regional levels has also occurred through the structures and provisions of Indigenous settlement agreements (e.g., Nunavut Act, 1993), fish and wildlife co-management agreements, and through various boundary organizations.. However, Indigenous involvement in several arenas remains limited, in part, because of insufficient financial resources to support participation (*high confidence*). While there has been great involvement of subsistence users in monitoring and research on climate change (see 3.5.5.1 below), resource management regimes that regulate harvesting are largely dictated by conventional, science-based paradigms that give limited legitimacy to the knowledge and suggested preferences of subsistence users (see Section 3.5.6.1 and Cross-Chapter Box 3 in Chapter 1).

The social costs and social learning of responding to climate change are linked. Involvement in adaptive co-management comes with high transaction costs (e.g., greater demands on overburdened Indigenous leaders, added stress of communities living with limited resources). In some cases, co-management can give communities greater voice, but can also perpetuate dominant paradigms of resource management (AMAP, 2018). The threat of climate change can at the same time reinforce cultural identify and motivate great political involvement, which turn, gives Indigenous leaders experience as agents of change in policy making. Penn et al. (2016) pointed to these conflicting issues, arguing the need for a greater focus on community capacity and cumulative effects.

3.5.3.5 Reindeer Herding

Herders’ responses to climate change have varied by region and respective herding practices, and in some cases constrained by limited access to pastures (Klokov, 2012; Forbes et al., 2016; Uboni et al., 2016; Mallory and Boyce, 2017) These conditions are exacerbated in some cases by high numbers of predators (Lavrillier and Gabyshev, 2018). In Fennoscandia husbandry practices of reindeer by some (mostly Sami) include supplemental feeding, which provide a buffer for unfavourable conditions. In Alaska, reindeer herding is primarily free range, where herders must manage herd movements in the event of icing events and the potential loss of reindeer because movements of caribou herds (wild reindeer), both of which are partially driven by climate. For Nenets of the Yamal, resilience in herding has been facilitated through herders’ own agency and, to some extent, the willingness of the gas industry on the Yamal to observe non-binding guidelines that provide for herders’ continued use of traditional migrations routes (Forbes et al., 2015). In response to climate change (i.e., icing events and early spring run offs blocking migration), the only way of avoiding high deer mortality is to change the migration routes or take the deer to other pastures. In practice, however, the full set of challenges has meant more Yamal herders opting out of the traditional collective migration partially or entirely to manage their herds privately. The reason to have private herds is one of adaptive advantage; smaller, privately-owned herds are nimbler in the face of rapid changes in land cover and the expansion of infrastructure (Forbes, 2013). The same logic has more recently been applied by some herders in the wake of recent rain-on-snow events (see Section 3.4.3.2.2) (Forbes et al., 2016). In all these regions, land-use changes that restrict movement of reindeer to pastures will negatively interact with the effects of climate, and affect the future sustainability of herding systems (*high confidence*).

3.5.3.6 Tourism

The growth of the polar tourism market is, in part, a response to climate change, as travellers seek 'last-chance' opportunities, which, in turn, is creating new challenges in governance. The anticipated increase in the near- and long-term future, especially with the travel of small vessels (yachts) (Johnston et al., 2017), points to a deficiency in current regulations and policies adequate to address human safety, environmental risks, and culture impacts. Polar-class expedition cruise vessels are now, for the first time, being purposefully built for recreational Arctic sea travel. Opportunities for tourism vessels to contribute to international research activities ('ships of opportunity'), may improve sovereignty claims in some regions, contribute to science, and enhance education among public about Arctic regions (Stewart et al., 2013; Arctic Council, 2015; Stewart et al., 2015). The anticipated grow of cruise tourism in polar regions also points to the need for operators, governments, destination communities, and others to identify and evaluate adaption strategies, such as disaster relief management plans, updated navigation technologies for vessels, codes of conduct for visitors, and improved maps (Dawson et al., 2016). As well, limited research has examined perceptions of tourism and appropriate adaptation responses by residents of local community destinations (Kaján, 2014; Stokke and Haukeland, 2017). Efforts were initiated with stakeholders in Arctic Canada to identify strategies that would lower risks. A next step in lowering risks and building resilience is to further develop those strategies (Dawson et al., 2016) (*medium confidence*).

Tourism activities in the Antarctic are coordinated by the International Association of Antarctic Tour Operators, which has worked with Antarctic Treaty Consultative Parties to manage changes in operations and their impact on ice-free areas (ATCM, 2016). It has been suggested that use of existing protected area management mechanisms be used to mitigate some of the impacts of high visitation rates (ASOC, 2015). However, there is a general disagreement about the regulation of Antarctic tourism among Treaty Parties and the benefits parties derive from tourism are currently not shared. Climate change is a challenge because it is often considered as an external factor that can be dealt with from a scientific perspective. Legal basis applying are the Madrid protocol (Art. 3) requiring a minimization of adverse environmental impacts vs. global environmental regimes (such as ATS) to a greater extent (Dodds, 2010; Hemmings and Kriwoken, 2010; Orheim et al., 2011; Triggs, 2011) (*medium confidence*).

3.5.3.7 Infrastructure

Reducing and avoiding the impacts of climate change on infrastructure will require special attention to engineering, land-use planning, and private and public budgeting. In the case of some communities, relocation will be required, necessitating more formal methods of assessing relocation needs and identifying sources of funding to support relocations.

Regional-to-local-level adaptation is affected by budgetary constraints, and maintenance of infrastructure (e.g., road maintenance) has already increased operating costs for local and regional governments. Melvin et al. (2017) estimated costs damages (without climate change adaptation measures) to public infrastructure in Alaska from 2015 to 2099 will be $4.2billion to $5.5 billion, depending on the climate scenario (Figure 3.12). Estimates of proactive adaptation measures (i.e., reduction of greenhouse gases) is estimated to reduce damage costs by half, or $1.4 billion. The analysis of Melvin et al. (2017) does not include the cost of damages to private infrastructure and is therefore, only a percentage of total costs. Because of the high number of large settlements and the dependence of the Russian economy on sectors based in northern regions, a triage infrastructure assessment approach may be helpful in that region.

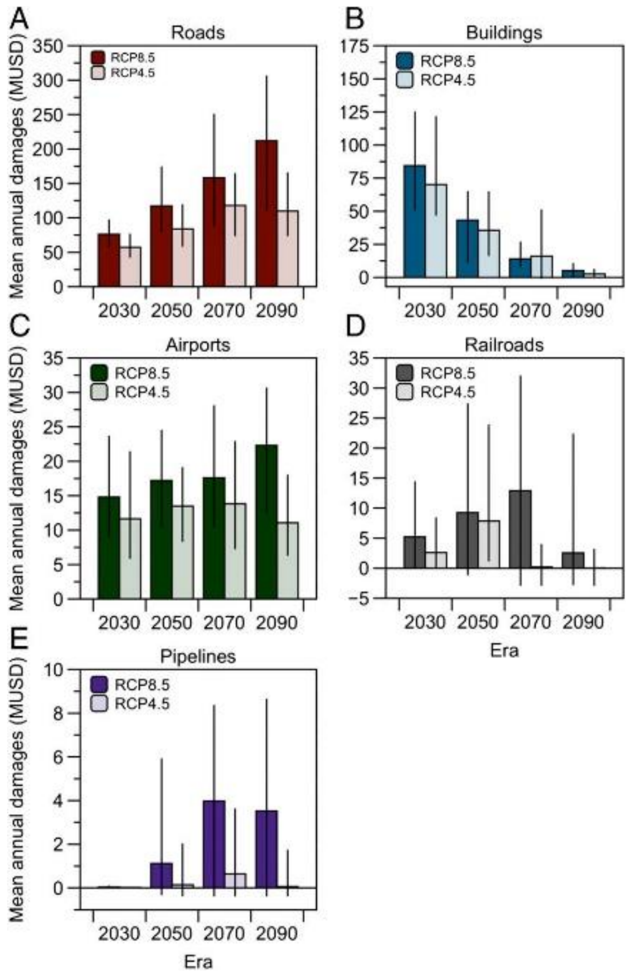


Figure 3.12: Annual damages [undiscounted and without adaptation in million US dollars (MUSD)] to each infrastructure type [(A) roads, (B) buildings, (C) airports, (D) railroads, and (E) pipelines] for four study eras. Values are the mean ± minimum, maximum for five GCMs and represent the mean annual damages (sum of all evaluated environmental stressors for each infrastructure type) for the 20 y included in each era. Note the difference in scales among panels. From Melvin et al. (2017) [PLACEHOLDER FOR THE FINAL DRAFT: Figure to be redrawn]

A discussion of the relocation of Alaska's coastal villages is found in Cross-Chapter Box 7. 12 Alaskan coastal communities are not, however, the only villages potentially requiring relocation. Subsidence due to thawing permafrost and river erosion makes other rural communities of Alaska and elsewhere vulnerable, and potentially requiring relocation in the future. The situation in Alaska raises issues of environmental justice and human rights (Bronen, 2017) and illustrates the limits of incremental adaptation when transformation change is needed (Kates et al., 2012). As noted, empirical evidence does not show an outmigration from Alaskan rural villages threated by climate change (Hamilton et al., 2016b). Huntington et al. (2018) point to people's attachment to place, their inability to relocate, the effectiveness of alternative ways of achieving acceptable outcomes, and methods of buffering through subsidies as explanations for limited outmigration responses and the desire to stay put.

3.5.3.8 Arctic Human Health and Well Being

Human health-related responses to climate change transcend multiple levels, from the individual to the international, and a number of initiatives are already underway to address them. In the future, the ability to manage, respond, and adapt to climate-related health challenges will be a defining issue for health sector in the polar regions (Blashki et al., 2011; Cunsolo Willox et al., 2012; Sibbald, 2013) (*high confidence*).

At present health adaptation to climate change is generally under-represented in policies, planning, and programming. For instance, all initiatives of the Fifth National Communications of Annex I parties to the United Nations Framework Convention on Climate Change affect health vulnerability, however, only 15% of initiatives had an explicit human health component described (Lesnikowski et al., 2011). The Arctic is no exception to this global trend. Despite the substantial health risks associated with climate change in the polar regions, health adaptation responses remain sparse and piecemeal (Lesnikowski et al., 2011; Panic and Ford, 2013; Ford et al., 2014b; Loboda, 2014), with the health sector substantially under-represented in adaptation initiatives compared to other sectors (Pearce et al., 2011; Ford et al., 2014b; National Research Council, 2015). Furthermore, the geographic distribution of publically available documentation on adaptation initiatives is skewed in the Arctic, with more than three-quarters coming from Canada and USA (Ford et al., 2014b; Loboda, 2014) (*medium confidence*).

Many health adaptation efforts by governments have been groundwork actions, focused increasing awareness of the health impacts of climate change and conducting vulnerability assessments (Lesnikowski et al., 2011; Panic and Ford, 2013; Austin et al., 2015). For instance, in Canada, this effort has included training, information resources, frameworks, general outreach and education, and dissemination of information to decision makers (Austin et al., 2015). Finland's federal adaptation strategy outlined various anticipatory and reactive measures for numerous sectors, including health (Gagnon-Lebrun and Agrawala, 2007). In Alaska, the Arctic Investigations Program responds to infectious disease via advancing molecular diagnostics, integrating data from electronic health records and environmental observing networks, as well as improving access to in-home water and sanitation services. Furthermore, circumpolar efforts are also underway, including an circumpolar working group with experts from public health to assess climate-sensitive infectious diseases, and to identify initiatives that reduce the risks of disease (Parkinson et al., 2014). Importantly, health adaptation is occurring at the local scale in polar regions (Ford et al., 2014a; Ford et al., 2014b). Adaptation at the local scale is broad, from community freezers to increase food security, to community-based monitoring programs to detect and respond to climate-health events, to Elders mentoring youth in cultural activities to promote mental health when people are 'stuck' in the communities due to unsafe travel conditions (Pearce et al., 2010; Harper et al., 2012; Douglas et al., 2014; Austin et al., 2015; Bunce et al., 2016; Cunsolo Willox et al., 2017) (*high confidence*). Several regional and national-level initiatives on food security (ICC, 2012), as well as research reporting high levels of household food insecurity (Kofinas et al., 2016; Watts et al., 2017) have prompted greater concerns for climate change (Loring et al., 2013; Beaumier et al., 2015; Islam and Berkes, 2016). At the international level, the Arctic Council launched the 'One Health' initiative, an effort to advance understanding of

One Health is a particularly well-matched tool to advance the understanding of health threats from the direct and indirect impacts of climate change in the Arctic. As a multidisciplinary approach, One Health strengthens coordination between and among a wide range of scientific disciplines and stakeholder. One Health enhances participatory community-based approaches for identifying and responding to health issues in communities which take into account Indigenous Knowledge and local knowledge.

Table 3.7: Human response of key systems / sectors to climate change in polar regions

System / Sector	*Consequence of climate change*	*Documented responses*	*Key assets and strategies of adaptive and transformative capacity*	*Anticipated future conditions / level of certainty*	*Other forces for change that may interact with climate and affect outcomes.*
Commercial Fisheries	Consequences are multi-dimensional, including impacts to abundance and distribution of different target species differently, by region. Changes in coastal ecosystems affecting fisheries productivity	Implementation of adaptive management practices to assess stocks, change allocations as needed, and address issues of equity	Implementation of adaptive management that is closely linked to monitoring, research, and public participation in decisions	Displacement of fishing effort will impact fishing operations in the CAMLR Convention area (*medium confidence*)	Changes in human preference, demand, and markets, changes in gear, changes in policies affecting property rights.
Subsistence (marine and terrestrial)	Changes in species distribution and abundance(not all negative); impediments to access of harvesting areas; safety; changes in seasonality; reduced harvesting success and process of food production (processing, food storage; quality); threats to culture and food security	Change in gear, timing of hunting, species switching; mobilization to be involved in political action	Systems of adaptive co-management that allow for species switching, changes in harvesting methods and timing, secure harvesting rights.	Less access to some areas, more in others. Changes in distribution and abundance of resources. More restrictions with regulations related to species at risk. Adaptation at the individual, household, and community levels may be serious restricted by conditions where there is poverty. *(high confidence*)	Changes in cost of fuel, land use affecting access, food preferences, harvesting rights; international agreements to protect vulnerable species.
Reindeer Herding	Rain-on-snow events causing high mortality of herds; shrubification of tundra pasture lowering forage quality	Changes in movement patterns of herders; policies to insure free – range movements.	Flexibility in movement to respond to changes in pastures, secure land use rights and adaptive management. Continued economic viability and cultural tradition.	Increased frequency of extreme events and changing forage quality adding to vulnerabilities of reindeer and herders (*medium confidence*)	Change in market value of meat; overgrazing; Land-use policies affecting access to pasture and migration routes, property rights.
Non-Renewable Resource Extraction (Arctic only)	Reduced sea ice and glaciers offering some new opportunities for development; changes in hydrology (spring runoff), thawing permafrost, and temperature affect	Some shifts in practices, greater interest in off shore and on-land development opportunities in coms regions.	Modification of practices and use of climate change scenario analysis.	Increased cost of operations in areas of permafrost thawing; more accessible areas in open waters and receding glaciers.	Changes in policies affecting extent of sea & land use area, new extraction technologies (e.g., fracking), changes in markets (e.g., price of barrel of oil)

	production levels, ice roads, flooding events, and infrastructure				
Transportation	Open seas allowing for more vessels; greater constraints in use of ice roads	Increased shipping, tourism, more private vessels. Increased risk of hazardous waste and oil spills and accidents requiring search and rescue.	Strong international cooperation leading to agreed upon and enforced policies that maintain standards for safety; well-developed response plans with readiness by agents in some regions	Continued increases in shipping traffic with increased risks of accidents. Shortening windows of operation for use of ice roads.	Political conflict in other areas that impeded acceptance of policies for safety requirements, timing, and movements. Changing insurance premiums.
Infrastructure -urban and rural human settlements, year-round	Thawing permafrost affecting stability of ground; coastal erosion,	Damaged and loss of infrastructure, increase in operating costs.	Resources for assessments, mitigation, and where needed, relocation.	Increasing cost to maintain infrastructure and greater demand for technological solutions to mitigate issues.	Weak regional and national economies, other disasters that divert resources, disinterest by southern-based law makers
Coastal settlements (See Cross-Chapter Box 7: Low-lying Islands and Coasts)	Change in extent of sea ice with more storm surges, thawing of permafrost, and coastal erosion	Maintenance of erosion mitigation; relocation planning, some but incomplete allocation for funding	Local leadership and community initiatives to initiate and drive processes, responsive agencies, established processes for assessments and planning, geographic options.	Increasing number of communities needing relocation, rising costs for mitigating erosion issues.	Limitations of government budgets, other disasters that may take priority for spending, deficiencies in policies for addressing mitigation and relocation
Tourism (Arctic and Antarctic)	Warmer conditions, more open water, Public perception of 'last chance' opportunities	Increased visitation, increase in off-season tourism to polar regions	Policies to insure safety, cultural integrity, ecological health, adequate quarantine procedures	Increased risk of introduction of alien species and direct effects of tourists on wildlife	Travel costs. Shifting tourism market, more enterprises
Human Health	Threats to food security, potential threats to physical and psychological well being	Greater focus on food security research; programs that address fundamental health issues	Human and financial resources to support public programs in hinterland regions; cultural awareness of health issues as related to climate change.	Greater likelihood of illnesses, food insecurity, cost of health care.	A reduction (of increase) in public resources to support health services to rural community populations, research that links ecological change to human health

3.5.4 Governance

This section describes and assesses the role of governance in human responses to climate change (see SROCC Annex I: Glossary for definition of governance). The role of governance in human responses is examined below with two categories: 1) local to national and 2) international. Governance in Polar Regions today functions both within and across levels of interaction as a web of multi-level transactions, through formal and informal institutions and social networks. Multilevel governance, as analytical tool, accounts for overlapping competencies and deficiencies among different levels that are vertical (from global, international, regional, national and local) and horizontal (from local to local, regional to regional). Interactions between difference levels and across various actors is in some cases are polycentric (see Cross-Chapter Box 1 in Chapter 1), and thus contribute to adaptability and resilience of social-ecological systems to climate change in Polar Regions. This assessment of climate change governance examines initiatives and evidence of cooperation and competition at and across multilevel governance with important examples albeit not in an exclusive fashion.

3.5.4.1 Local-to National Governance

Responses to climate change at and across local, regional, and national levels occurs directly and indirectly through a broad range of governance activities, such as land- and sea-use planning and regulations, economic development strategies, tax incentives for use of alternative energy technologies, permitting processes, resource management, and national security. Increasingly climate change is considered in environmental assessments and proposals for resource planning of polar regions.

Ford et al. (2014b) comprehensive literature review of 157 discrete cases of arctic adaptation initiatives found that adaptation is primarily local, and motivated by reducing risks and their related vulnerabilities. Several elements for successful climate change adaptation planning at the local level have previously been identified: formal analytical models need to be relevant to the concerns and needs of stakeholders, ‘experts’ be made aware of and sensitive to community perspectives, information should be packaged and communicated in ways that is accessible to non-experts, and processes of engagement that foster creative problem solving be used (Sheppard et al., 2011). Success of local government involvement in adaptation planning has been closely linked to transnational municipal networks which foster social learning and in which local governments assume a role as key players (Fünfgeld, 2015). Fünfgeld (2015) also noted that while evidence shows that transnational networks can be a catalyst for action and promoting innovation, there remain outstanding challenges in measuring the effectiveness of these networks.

Adaptation through formal institutions by Indigenous People is potentially enabled through self-government, land claims, and co-management institutions (Baird et al., 2016; Huet et al., 2017). However, as studies have found organizational capacity to be a limiting factor in involvement (Ford et al., 2014b; AHDR, 2015; Forbes et al., 2015). The interactions across scales is also dependent on the extent to which various stakeholders are perceived as legitimate in their perceptions and recommendations, an issue related to the use of traditional knowledge in governance (see Cross-Chapter Box 3 in Chapter 1). Interestingly, Cashmore and Wejs (2014) study of local-level climate change planning found that moral and ethical reasoning had low salience as compared to cultural legitimacy in their case studies.

At a more regional level, Alaska’s ‘Climate Action for Alaska’ was reconstituted in 2017 from an initiative previously launched by Governor Palin and is now actively linking local concerns with state-level policies and funding, as well as setting targets for future reductions in the state’s carbon-emission. Research in Norway showed the important role of these cross-scale boundary organizations have in climate change adaptation planning, and how central government initiatives can ultimately translate into ‘hybrid’ forms of adaptation at the local level that allow for actions that are sensitive to local communities (Dannevig and Aall, 2015).

At the national level, Norway, Sweden, and Finland have engaged in the European Climate Adaptation Platform (‘Climate-ADAPT’), a partnership between the European Commission and the European Environment Agency. Climate-ADAPT aims to support Europe in adapting to climate change. As an initiative of the European Commission and helps users to access and share data and information on expected climate change in Europe, current and future vulnerability of regions and sectors, national and transnational adaptation strategies and actions, adaptation case studies and potential adaptation options, and tools that support adaptation planning. Level of participation by country and the extent to which national level efforts are linked with regional and local adaptation varies. In spite of the US government's withdrawal from the Paris Agreement, the US is now completing it second National Climate Change Assessment, which includes a focus on Alaska's Arctic and subarctic regions. The Canadian government's actions on climate change have been the most extensive of the Arctic nations, including development of the Arctic Climate Portal, its lead role in the Arctic Adaptation Climate Assessment process which generated three in-depth regional reports, and consideration of climate change by The Northern Contaminants and Nutrition North Canada programs.

3.5.4.2 International Climate Governance and Law in Polar Regions: Implications for International Cooperation

Responses to climate change in international cooperation are assessed within different levels of governance, and with different institutional arrangements. Such arrangements involve both formal and informal actors, as well as networks operating with different norms (see also Cross-Chapter Box 2 in Chapter 1). Below and in the following sections such legal frameworks and the role of environmental institutions in managing multilevel governance is assessed, highlighting the fragmentation of international law and the synergistic linkages and reverberations among different levels and sectors of governance.

The way states and institutions manage international cooperation on environmental governance is changing in response to climate change in the polar regions. Rather than treating regional impacts of climate change and their governance in isolation (i.e., purely with a regional lens), the need to cooperate in a global multi-regulatory fashion across several levels of governance is increasingly realised (Stokke, 2009; Cassotta et al., 2016; Keil and Knecht, 2017) (*medium confidence*).

In both polar regions, innovative cooperative approaches to regional governance are developed that allow for the participation of non-state actors. In some cases, inclusive decision-making procedures of these regimes allow for a substantial level of participation by specific groups of the civil society, such as stakeholders. For example, in the Antarctic Treaty System (ATS), the parties to the Convention for the Conservation of Antarctic Seals of 1972 have granted a central role in its regime to the Scientific Committee on Antarctic Research (SCAR), a non-governmental scientific organization. In the Arctic, the status of Permanent Participants has enabled the effective participation of Indigenous Peoples in the work of the Council (Pincus and Ali, 2016). Climate change contributes to modifying the balance between the interests of official and non-official actors, leading to changing forms of cooperation (Young, 2016). While such changes and modifications occur in both the Arctic and Antarctic, the role of states has remained present in all the regimes and sectors of human responses (Young, 2016; Jabour, 2017).

Addressing the risks of climate change impacts in polar regions also requires linking levels of governance and sector governance across local to global scales, considering impacts and human adaptation (Stokke, 2009; Berkman and Vylegzhanin, 2010; Tuori, 2011; Young, 2011; Koivurova, 2013; Prior, 2013; Shibata, 2015; Young, 2016) (*high confidence*). Despite established cooperation in international polar region governance, several authors come to the conclusion that the current international legal framework seems inadequate to manage in a precautionary approach (*medium confidence*). For example, several studies have shown that the Convention on the Protection of the Marine Environment of the North East Atlantic (OSPAR) that provides a framework for implementation of UNCLOS and the Convention on Biological Diversity (CBD), are insufficient to deal with risks applying a precautionary approach (Jakobsen, 2014; Hossain, 2015).

In the Arctic, responses to climate change do not only lead to international governance cooperation but also to competition in access to natural resources, especially fossil fuel. With ice retreating and thinning, leading to easy access to natural resources, coastal states are increasingly using Art. 76 of the UNCLOS (Art. 76 UNCLOS; Verschuuren (2013)), which relates to the extension of territorial jurisdiction, which states would acquire once they can demonstrate with scientific data that their continental shelf is extended. In that case they can enjoy sovereign rights beyond the Exclusive Economic Zone (EEZ). It is *very unlikely* that this new trend from states to refer to Art.76 will lead to future (military) conflicts (Berkman and Vylegzhanin, 2013;

Kullerud et al., 2013; Stokke, 2013; Verschuuren, 2013), although the issue cannot be totally dismissed (Kraska, 2011; Åtland, 2013; Huebert, 2013; Cassotta et al., 2015; Barret, 2016; Cassotta et al., 2016).

In the Antarctic, the variety of economically viable resources is limited. At present, the focus is on the only two economically viable resources: marine living resources and tourism. Currently cooperation does occur via UNCLOS, the Convention for the Safety of Life at Sea (SOLAS) and the Convention for the Prevention of Pollution from Ships (MARPOL) and the Polar Code, which applies to tourism vessels, and through the International Association for Antarctic Tourism (IAATO) managing of tourism in accordance to the Antarctic Treaty System (ATS). Cooperation in the Antarctic also occurs with the CCAMLR falling under the ATS. Climate change is a central issue for the CCAMLR because it poses challenges regarding its impact on waters and the way to regulate and manage fisheries.

3.5.4.2.1 Formal arrangements: polar conventions and institutions

Both in the Arctic and the Antarctic, international cooperation in different sectors to identify responses alleviating cryosphere-related climate change impacts on people and ecosystems is well established (Vidas, 2007; Stokke, 2009; Berkman and Vylegzhanin, 2010; Wilson Rowe, 2013; Barret, 2016; Wehrmann, 2016; Young, 2016) (*high confidence*). Several instruments of cooperation that operate at the global level are the basis for vertical implementation to the polar oceans, such as the United Nations Convention on the Law of the Sea (UNCLOS), a global convention that codifies customary international law (Birnie et al., 2009; Dixon, 2013).

The Arctic Council

International cooperation on issues related to climate change in the Arctic mainly occurs at the Arctic Council (herein 'the Council'), and consequently in important areas of its mandate: the (marine) environment and scientific research (Koivurova, 2016; Tesar et al., 2016; Wehrmann, 2016; Young, 2016). The Council is an example of cooperation through soft law, a middle-way and unique meta-juridical institutional body. It is increasingly operating in a context of the Arctic affected by a changing climate, globalization and transnationalism (Baker and Yeager, 2015; Cassotta et al., 2015; Pincus and Speth, 2015) (*medium confidence*). In 2013, the Council granted China, South Korea, Japan, India, Italy and Singapore the status of observers at the Council.

Despite lacking the role to enact hard law, the Council undertook the signature of three binding agreements, the latest of which is the Agreement on Enhancing International Arctic Scientific Cooperation, which is an indication the Council is preparing a regulatory role to responding to climate change in the Arctic using hard-law instruments (Koivurova, 2016; Shapovalova, 2016). Through organising the Task Force on Black Carbon and Methane (Koivurova, 2017), the Council has catalysed action on short-lived climate forcers as the task force was followed by the adoption in 2015 of the Arctic Council Framework for Action on Enhanced Black Carbon and Methane Emission Reductions. In this non-legally binding agreement, Arctic States lay out a common vision for national and collective action to accelerate decline in black carbon and methane emissions (Shapovalova, 2016). The Council thereby moved from merely assessing problems to attempting to solve them (Baker and Yeager, 2015; Young, 2016; Koivurova and Caddell, 2018). While mitigation of global emissions from fossil fuels require global cooperation, progress with anthropogenic emissions of short-term climate forcers (such as black carbon and methane) may be achieved through smaller groups of countries (Aakre et al., 2018). However, even though the Council has also embraced the concept of ecosystem management, it does not have working groups with a mandate to address fisheries issues related to climate change applying a precautionary approach.

Several studies have shown that the Council has the potential to enhance internal coherence in the current, fragmented landscape of multi-regulatory governance by providing integrated leadership. However, it is about *as likely as not* that the Council could play a strong role in combatting global climate problems and operating successfully within the climate transnational context unless going through restructuring and reconfiguration (Stokke, 2013; Baker and Yeager, 2015; Pincus and Speth, 2015; Cassotta et al., 2016; Tesar et al., 2016; Wehrmann, 2016; Young, 2016; Koivurova and Caddell, 2018).

The future of the governance of the changing Arctic Ocean, including the role of the Council will also depend on the implications of the recent new agreement on the Conservation and Sustainable use of Marine Biodiversity of Areas beyond National Jurisdictions (BBNJ), signed in December 2017 under the United

Nations Convention on the Law of the Sea (UNCLOS) (Baker and Yeager, 2015; De Lucia, 2017; Nengye et al., 2017; Koivurova and Caddell, 2018) (*medium confidence*).

Governance involving Indigenous Peoples

Several organizations represent Indigenous interests at the international scale, are actively involved in climate change governance, and partake as Permanent Participants at the Arctic Council. Among those organizations, the Inuit Circumpolar Council, a non-governmental organization is representing about 160,000 Inuit living in four Arctic countries, is the most active, operating also at the UN level with special consultative status with the United Nations Economic and Social Council. The Inuit Circumpolar Council's role is noteworthy in raising Indigenous issues on climate change at the global level and supporting adaptation policies important to Inuit people. In supporting adaptation policies at the regional level, the Inuit Circumpolar Council has also worked in conjunction with the Council's working groups to highlight the need of investment in new infrastructure that assists Inuit communities' adaptation changes in climate, sea ice and shorelines and the move away from carbon fuels (Inuit Circumpolar Council Canada, 2014). Inuit Circumpolar Council along with other Indigenous organizations, do not have the right to vote at the Council, which limits its influence affecting the Council's resolutions and policies addressing new risks and uncertainties.

The Antarctic Treaty System (ATS)

The importance of understanding, mitigating and adapting to the impacts of changes to the Southern Ocean and Antarctic cryosphere has been realized by all of the major bodies responsible for governance in the Antarctic region (south of 60°S). The Antarctic Treaty Consultative Parties agreed that a Climate Change Response Work Programme would address these matters (ATCM, 2016). This led to the establishment of the Subsidiary Group of the Committee for Environmental Protection on Climate Change Response (ATCM, 2017), which recognized the importance of climate change in its area of interest. As its last meeting, however, the committee was unable to agree a Climate Change Response Work Program (CCAMLR, 2017a). Consensus on a work program will be needed to achieve tangible progress in this area.

3.5.4.2.2 Informal arrangements

Climate change in the Arctic is facilitating access to natural resources (see Section 3.5.3.3) which may generate financial capital for Arctic residents and their governments, including Indigenous Peoples. Indigenous Peoples are considered as non-state actors and in many cases promote environmental protection in support of the sustainability of their traditional livelihoods. This protection is at times in opposition to the pro-development business sector, which is well-funded and lobbies strongly. Bilateral agreements for resource development in the Arctic are typically state dominated and controlled, and are negotiated with powerful non-state actors, such as China National Petroleum Company; state-dominated companies such as Gazprom or Statoil and private corporations like Exxon Mobil (Young, 2016). Among the non-state actors, new networks and economic forums have been established (Wehrmann, 2016). One example is the Arctic Economic Council (AEC), created by the Council during 2013-15 as an independent organization that facilitates Arctic business-to-business activities and supports economic development.

The Antarctic Treaty Consultative Parties, through the Subsidiary Group of the Committee for Environmental Protection on Climate Change Response, continue to work closely with the Scientific Committee on Antarctic Research, the Council of Managers of National Antarctic Programs, the International Association of Antarctica Tour Operators and other NGOs to understand, mitigate and adapt to impacts associated with changes to the Southern Ocean and Antarctic cryosphere. Various bilateral and multi-lateral projects are underway to understand and mitigate risk, with many of these funded by national programs. Understanding, mitigating and adapting to climate change are among the key priorities identified for research in the region (Kennicutt et al., 2014a; Kennicutt et al., 2014b), and progress has been done to understand how best to support such work (Kennicutt et al., 2016) and ensure that its implications reach policy-makers (CEP, 2017).

3.5.4.2.3 Role of informal actors

Several studies show that informal actors of the Arctic can influence decision-making process of the Council and shift the Council towards more cooperation with distinct actors to enhance the co-production of knowledge (Duyck, 2011; Makki, 2012; Keil and Knecht, 2017). Recently, non-state observers at the Council, such as the World Wide Fund for Nature (WWF) and the Circumpolar Conservation Union (CCI) have played a role in raising awareness on climate change responses and contributing to the work of the Council's Working Groups and Expert Groups (Keil and Knecht, 2017). However, few studies have concentrated on the role of informal actors in Arctic governance (Duyck, 2011; Makki, 2012; Keil and Knecht, 2017).

Within the Antarctic Treaty System, several non-state actors play a major role in providing advice and influencing the governance of Antarctica and the Southern Ocean. Among the most prominent actors at the Antarctic Treaty Consultative Meetings are formal observers such as the Scientific Committee on Antarctic Research, and invited experts such as the International Association of Antarctica Tour Operators and the Antarctic and Southern Ocean Coalition. At meetings of the Convention on the Conservation of Antarctic Marine Living Resources, invited observers include organisations such as Antarctic and Southern Ocean Coalition, International Association of Antarctica Tour Operators and Scientific Committee on Antarctic Research, and representatives of industry such as the Association of Responsible Krill harvesting companies. The Scientific Committee's 2009 report on Antarctic Climate Change and the Environment (ACCE) (Turner et al., 2009) precipitated an Antarctic Treaty Meeting of Experts on Climate Change in 2010 (Antarctic Treaty Meeting of Experts, 2010). The outcomes of the meeting led the Antarctic Treaty's Committee for Environmental Protection (CEP) to develop a Climate Change Response Work Programme, which is now overseen by a formal Subsidiary Group on Climate Change Response (ATCM, 2017).

3.5.5 Towards Resilient Pathways

Resilient pathways (see SROCC Annex I: Glossary) are a necessary complement to climate adaptation. In this respect, adaptation is best viewed not as a destination, but one part of an on-going, iterative social learning process of responding to immediate climate change impacts while building the capacity of society to respond to likely and unknown future conditions and achieves its goals of sustainable development (see SROCC Annex I: Glossary) (AMAP, 2017a; AMAP, 2017b; AMAP, 2018), Resilience pathways are guided by seven actionable items noted in Cross-Chapter Box 1 in Chapter 1. Here we describe a select set of innovative practices currently utilized in polar regions that relate to those areas of action, and which have proven potential for building resilience in the face of climate change and its uncertainties. They include: Knowledge co-production and integration, with a focus on community-based monitoring; Scenario analysis and planning with a focus on participatory approaches; the linking of knowledge with action; and strategies for ecosystem stewardship. This list of practices is not all inclusive of the many innovative efforts for responding to climate change underway in polar regions, but together they represent interrelated elements of adaptive governance that operationalize resilience building. In some cases, the practices described are novel and in their nascent stages of development, and hence, require more refinement. Others are well developed. All, however, have shown sufficient utility to merit further use (ARR, 2016) (*high confidence*).

3.5.5.1 Knowledge Co-production and Integration

The challenges of climate change in polar regions require a new paradigm in knowledge production that moves beyond single disciplinary investigations to transdisciplinary approaches that benefit from the insights of a diversity of cultural, geographic, and disciplinary perspectives (Armitage et al., 2011; Johnson et al., 2015a; Rycroft-Malone et al., 2016; Berkes, 2017; Miller and Wyborn, 2018; Robards et al., 2018). Knowledge co-production for sustainability is most effective when it uses a social-ecological frame that problematizes phenomena to engage a broad set actors with diverse epistemological orientations on what is known and how it is known (Berkes, 2017). Team work and good leadership are critical elements of the knowledge co-production process (Meadow et al., 2015; National Research Council, 2015). Knowledge co-production as defined here includes observing (i.e., documenting observations, identifying key variables to monitor/ identification of indicators of change; data archiving), and the analysis of data to improve understanding causality, trends, or emergent patterns. The linking of knowledge with policy and action, also part of the process, is discussed below in Section 3.5.6.2 (e.g., US SEARCH Program).

Knowledge co-production addressing climate change is well suited for polar regions, given the cultural diversity of the Arctic and the international cooperation of Antarctica monitoring and research. It is currently employed to varying degrees at almost all levels and in many regions of the Poles.

A noteworthy activity \ in Arctic knowledge co-production has been in the implementation and development of innovative community-based monitoring initiatives (Lovecraft et al., 2013; Johnson et al., 2015a; Johnson et al., 2015b; Tomaselli et al., 2018). At the local level, communities, in many cases working in collaboration with agencies and academics, are documenting local observations of change, using narratives, semi-structured, new technologies such as camera-equipped GPS, and the phone apps. One example of innovation at the regional scale is the Local Environmental Observer network, which is using mobile phone technology and the internet to post observations, and scheduled phone conferencing to communicate and discuss unusual observations. ELOKA, a circumarctic initiative to address and disseminate methods (Pulsifer et al., 2012), programs through CAFF, and the recent release of the international Sustaining Arctic Observing Network Strategy and Implementation document are examples of grander scale Arctic efforts (Lee et al., 2015).

Although having great potential, executing community based monitoring has proven to be labour intensive and hard to sustain, requiring sufficient long-term financial support and human capital, and in some cases, the involvement of boundary organizations to provide support and bridge across the many levels of governance (Robards et al., 2018). As with all knowledge production process, power relationships (who decides, who is viewed as a legitimate knowledge holder, who gets access to resources for involvement, who benefits) underpin collaboration in Arctic and Antarctic knowledge production. One possible outcome in future efforts, therefore, is that community-based monitoring (and the contributions of local and traditional knowledge systems in general) may function in a separate sphere from more conventional science efforts, with little interaction. Another possible outcome is that efforts will be made to 'integrate' local knowledge observations with science in ways that discount locals' understanding of change, and with little benefit to local communities. While the practice of community-based monitoring has advanced recently, more development is needed to insure that it is well linked with research and policy making (*high confidence*).

3.5.5.2 Linking Knowledge Systems with Decision Making

Society is entering an era commonly branded as 'post-trust', or even 'post truth' (Lubchenco, 2017). Polling indicates that most people still believe that decision-making gains accuracy and legitimacy when science informs the process with objective evidence, but inherent tensions between science-based assessment and interest-based policy often prevent direct connectivity. Scientists and policy makers involved in areas of polar governance typically work in separate spheres of influence, tend to maintain different values, interests, concerns, responsibilities and perspectives, and gain minimal exposure to the other's knowledge system (see Liu et al., 2008). Information exchange also flows unequally, as officials struggle with information overload and proliferating institutional voices, while scientists perceive little feedback (Powledge, 2012). Further, the longstanding science mandate to remain 'policy neutral' typically leads to norms of constrained interaction. For these and other reasons, channels between the two camps often seem 'rudimentary at best' (Neff, 2009) (*medium confidence*).

There is a growing expectation in polar regions for a more deliberate strategy linking science with policy in an iterative process of regular interactions among scientists, resource managers, and stakeholders to enhance social learning about climate change and ways of responding. This redefined 'actionable science' can better support decisions by creating more rigorous and accessible products, while shaping a narrative that instils public confidence (Beier et al., 2015; Fleming and Pyenson, 2017). Participatory simulation modelling, structured decision making systems, visualization, and decision theaters are a few tools currently being developed to link science and policy in the polar regions (Schartmüller et al., 2015; Kofinas et al., 2016; Garrett et al., 2017; Holst-Andersen et al., 2017; Camus and Smit, 2018).

A key adjustment to business as usual, however, involves willingness to provide active decision-support services, more often than mere decision-support products (Beier et al., 2015). In short, circumstances call for a new breed of polar scientist who not only understands policy considerations, but also engages in policy formulation. Polar scientists can do much more to make their work widely available for use, including: enhanced data collaboration at every scale, more strategic social engagement, communication that both informs decisions and improves climate literacy, and explicit creation of consensus documents that provide interpretive guidance about research implications and alternative choices (Gewin, 2014). In many cases, successful efforts in linking science with policy follow from effective communication and personal relationships of trust (*high confidence*).

3.5.5.3 Scenario Analysis and Planning

Assessing future risk and responding to polar climate change in conditions of uncertainty will depend, in part, on methods for exploring plausible, likely and desirable futures with stakeholders, scientists, and policy makers (Resilience Alliance, 2010; ARR, 2016; Flynn et al., 2018). Participatory scenario analysis is a quickly evolving field of practice in polar regions and beyond, and has proven particularly useful for supporting climate adaptation when it engages stakeholders and uses a social-ecological perspective (ARR, 2016; AMAP, 2017a). Crépin et al. (2017) noted various nuanced system dynamics to be considered, such as long-fuse big bang processes, pathological dynamics, and unforeseen processes. Others have focused on process methods for engaging stakeholders in evaluating a system's resilience, adaptability, and transformability to both reduce risk and build resilience (Resilience Alliance, 2010). For example Blair et al. (2014) and others have noted the importance of accounting for stakeholders' perceptions of risk as part of assessments and scenario planning.

In the polar regions participatory scenario analysis has been applied to a variety of problem areas related to climate change impacts and with many approaches. Scenario planning in Antarctica is in its early stages of development (Liggett et al., 2017), although it has been applied in sub-Antarctic Chile to address marine spatial planning (Nahuelhual et al., 2017). The Canadian Department of National Defence used scenario analysis to study the national security issues of an ice-free Arctic and several workshops have drawn on scenarios to explore the implications of shipping in an ice-free arctic. In the Barents region, scenario workshops have included local and regional actors from public agencies, organizations and the private sector in three different locations to consider climate adaptation (AMAP, 2017b). In Alaska, the National Park Service's scenarios program, 'Rehearsing the Future', was used it to address possible futures, including climate change, with stakeholders (Ernst and van Riemsdijk, 2013). And at a more local scale, reindeer-herding youth across the Eurasian Arctic explored possible futures of climate change with other forces for change (van Oort et al., 2015; Nilsson et al., 2017).

Agencies and institutes have produced information and tools to support scenario planning (AMAP, 2015). The Scenarios Network for Alaska and Arctic Planning (SNAP) downscales GCMs to the local community scale, communicates data and outputs in user friendly formats, and engages of stakeholders through partnership programs (see https://www.snap.uaf.edu). The Oil Development Scenarios Project of the North Slope Science Initiative of Alaska (Vargas-Moreno et al., 2016) used maps in a participatory process that led to the identification of research needs. Flynn's (Flynn et al., 2018) review of scenario analysis in the Arctic, however, found that while the practice is widespread, less than half scenarios analyses incorporated climate projections. Flynn et al. (2018) also found that most studies utilizing a forecasting approach along with a backcasting approach had higher local participation, and that integrating different knowledge systems and cultural factors may have a higher impact on the utility and acceptance of the approach.

Clearly, analytical and the participatory approaches to scenario analysis have potential to enhancing knowledge co-production, and informing decisions on adaptation and building social-ecological resilience. The long-term utility of this practice will depend on the science of climate projections, further development of decision support systems to inform decision makers, as well as refinement of processes that facilitate stakeholder dialogue (*medium confidence*).

3.5.5.4 Resilience-based Ecosystem Stewardship

Resilience-based ecosystem stewardship, by definition, differs from conventional resource management or integrated ecosystem management, while retaining many of the principles of those two paradigms (Chapin III et al., 2009; Chapin III et al., 2010) (Table 3.8) In the polar regions, stewardship of ecosystems requires a focus on trajectories of change (i.e., emergence), implying that maintaining ecosystems in a state of equilibrium is not possible (Biggs et al., 2012; ARR, 2016).

Several stewardship strategies can reduce impacts and risks to species, habitats and ecosystems in support of social-ecological resilience. The first implements the tools of biodiversity conservation. Often expressed to protect the intrinsic values of biodiversity, they are increasingly understood as also supporting sustainable use of the environment (Ban et al., 2014), securing options for livelihoods (Salafsky and Wollenberg, 2000), and facilitating biodiversity adaptation in a changing environment (Mawdsley et al., 2009). In particular, networks of protected areas (vs isolated protected areas) are conceptualised (McLeod et al., 2009), planned (Solovyev et al., 2017) and implemented (Juvonen and Kuhmonen, 2013) to protect ecologically connected tracts of representative habitats, and biologically and ecologically significant features. While individual protected areas may prove problematic in a rapidly changing ecosystem, protected area networks that combine both spatially rigid and spatially flexible regimes with climate refugia operate in support of ecological resilience to climate change by maintaining genetic connectivity and flows, reducing direct pressures on biodiversity, and thus, giving biological communities, populations, and ecosystems the space to adapt (Nyström and Folke, 2001; Hope et al., 2013; Thomas and Gillingham, 2015). The second strategy is to maintain a continued flow of ecosystem services, while recognizing how of their benefits provide incentives for preserving biodiversity while ensuring options for sustainable development (Guerry et al., 2015). Incorporating an account of polar region ecosystem services into policy is a key method for integrating environmental, economic, and social policies that build resilience (CAFF, 2015). Currently, there is limited recognition of the wide range of benefits people receive from polar ecosystems, and a lack of planning and management tools that can demonstrate their benefits in decision-making processes (CAFF, 2015).

At national and international scales, two stewardship strategies have emerged in polar regions. One is to reduce global pressures that drive arctic climate change by reducing rates of greenhouse gas emissions. The second is to reconcile and coordinate local, regional, and national conservation actions through adaptive co-management, boundary organizations, visionary leadership and social networks. Opportunities for Arctic stewardship at landscape, seascape, and community scales, to a great extent, lie in supporting culturally engrained (often traditional Indigenous) values of respect for nature, and reliance on the local environment through the sharing of knowledge and power between local users of renewable resources and agencies responsible for managing these resources (Mengerink et al., 2017). In the Antarctic, ecosystem stewardship is highly dependent on international formally defined and informally enacted cooperation (*high confidence*).

Further implementation of ecosystem stewardship in polar regions will be driven, in part, by shifts in values about the value of biodiversity, a greater awareness of impacts of climate change in no action scenarios, and the political will to enact and support stewardship for the long-term benefit of society.

Table 3.8: Differences between conservation approaches focused on species, landscapes, and the mutual wellbeing of people and nature. (Chapin III et al., 2015)

Characteristic	**Species Conservation**	**Landscape and Seascape Conservation**	**Stewardship**
Reference point	Historic condition	Historic and current condition	Pathways of change
Central goal	Species protection	Conservation of ecosystem structure and function to conserve biodiversity and the habitats that support it	Sustain social-ecological systems and resilience of ecosystem services by fostering diversity
Predominant approach	Maintain species, populations, and habitats	Integrated management of human activities in landscapes and seascapes	Manage stabilizing and amplifying feedbacks
Role of protected areas	Habitat that is relatively safe from direct human impacts	Part of the habitat mosaic that interacts with unprotected habitat	Part of a complex social-ecological system that supports conservation and interacts with other societal goals
Role of uncertainty	Reduce uncertainty before taking action	Reduce uncertainty yet act in its presence	Embrace uncertainty: Maximize flexibility to adapt to an uncertain future
Role of resource manager(s)	Decision maker who sets course for sustainable management of species, populations, and habitats	Decision maker who sets course for sustainable management of landscapes, seascapes, and their components	Coordinated facilitators at multiple scales who engage stakeholder groups to respond to and shape social-ecological change and nurture resilience

Response to disturbance	Minimize disturbance probability and impacts	Minimize disturbance probability and impacts	Disturbance cycles used to provide windows of opportunity
Resources of primary concern	Species, populations, and habitats	Species, populations, and landscapes or seascapes	Biodiversity, human wellbeing, and adaptive capacity

3.5.6 Conclusion

Our assessment of human responses to climate change in polar regions reveals that all sectors of polar social-ecological systems are responding to the effects of climate change. The responses range, from having to incur an increase in operation costs (oil and gas industry, cost of government to maintain public infrastructure) to tourism operators taking advantage of emerging opportunities and new markets, harvesters of wild foods and reindeer herders modifying traditional practices while being exposed to greater risk, in the most extreme cases of communities planning to relocate settlements with limited support of public funds and reindeer herders abandoning their traditional livelihood. The more promising findings of this assessment relate to the development of resilience pathways for polar systems, and the need for continued and increased levels of cooperation and innovation in areas of knowledge co-production and multi-level governance that link local-to-global interactions in two-way vertical and horizontal directions. While promising, the degree to which so many sectors must respond speaks to the significant need for all actors of the polar regions to experiment, refine strategies, tools, and institutions that support on-going social learning *(high confidence)*.

3.6 Key Knowledge Gaps and Uncertainties

This chapter has assessed the state of knowledge concerning climate change impacts in the Polar Regions, their global influences, risks and potential responses. Progress will require that future assessments demonstrate increased confidence in various aspects; this can be achieved by closing numerous gaps in knowledge. Some of the key ones, which are priorities for future initiatives, are outlined here.

There is currently inadequate understanding of the mechanisms that have determined the observed changes and trends in Antarctic sea ice, notably the decadal increase in extent followed by the very recent rapid retreat. This has consequences for climate, ecosystems and fisheries, however the current lack of mechanistic understanding and poor model performance in reproducing observations translates to there being very limited predictive skill.

Overturning circulation in the Southern Ocean a key factor that controls heat and carbon exchanges with the atmosphere, and hence global climate, however there are no direct measures of this and only sparse indirect indicators of how it may be changing. This is a critical weakness in sustained observations of the global ocean.

Snow depth on sea ice is essentially unmeasured, limiting mass balance estimates and ice thickness retrievals. There is also large inconsistencies in trends of snow water equivalent (SWE) over Arctic land areas, reducing confidence in the assessments of snow's role in the water cycle and in insulating the underlying permafrost. Understanding of precipitation in the polar regions is critically limited by sparse observations, and there is a lack of understanding of the processes that drive regional variability in wetting/drying and greening/browning of the Arctic land surface.

There are clear regional gaps in ecosystems knowledge in the polar regions, and insufficient population estimates and reliable trends for many key species. Concerning assessments of ecosystems status, there are key uncertainties regarding the potential for organisms in the polar regions to adapt physiologically and behaviourally to habitat change, and also regarding the resilience of foodweb structures and the implications of changes in them for energy flow in the polar regions. Relatedly, knowledge gaps exist concerning how fisheries target levels will change alongside environmental change and how to incorporate this into decision making. Similarly, there are knowledge gaps on the extent to which changes in the availability of resources to subsistence harvesters affects food security of households.

Concerning polar glaciers and ice sheets, and the need to better understand their evolution and impacts on global sea level, longer quantifications of their changes are required, especially in regions where mass losses are greatest, and (relatedly) better attribution of natural versus anthropogenic forcings of these. Understanding of whether recent changes in West Antarctica represents the onset of an irreversible process is critically needed, as is understanding of the extent to which East Antarctica is sensitive to marine ice sheet instability. The response of subglacial hydrological systems to climate change requires improved understanding, as do the potential feedbacks to ice dynamics and ice sheet mass balance.

In terms of responding to climate change, reducing vulnerability and strengthening resilience, there is uncertainty on the effectiveness and limits of known adaptation strategies for reducing risk for people and sustainable development vis-a-vis the speed and complexity of projected impacts. In particular, while the occurrence of regime shifts in polar systems is both documented and anticipated, there is little or no understanding of their preconditions or of indicators that would help pre-empt them. Underlying this limited understanding lies the current considerable knowledge gap on how to link existing theoretical understandings of social-ecological resilience to practice in decision making and governance. At the practical level, there is also limited understanding concerning the resources that are needed to action successful adaptation responses. At the institutional level, key knowledge gaps exist about their effectiveness in supporting adaptation.

[START FAQ3.1 HERE]

FAQ 3.1: How do changes in the Polar Regions affect other parts of the world?

Climate change in the polar regions affect people in other parts of the world in two key ways. First, socio-economic impacts of ecosystem and physical changes in the polar regions extend across the globe. Second, physical changes in the Arctic and Antarctic alter processes that are important for global climate and sea level.

Among the impacts on societies and economies, aspects of food provision, transport, and access to non-renewable resources are of great importance. Fisheries in the polar oceans support regional and global food security, and are important for the economies of many countries around the world. Climate change alters Arctic and Antarctic marine habitats, and the ability of polar species and ecosystems to withstand or adapt to physical changes. This has consequences for fisheries production in the polar regions. Impacts will vary between regions, depending on the degree of climate change and the effectiveness of human responses. While management in some polar fisheries is among the world's most developed, expanding the implementation of precautionary, ecosystem-based, integrated, and adaptable governance will lower the impacts of climate change on marine ecosystems and fisheries.

New maritime shipping routes through the Arctic offer significant trade benefits because they are shorter than traditional passages via the Suez or Panama Canals. Shipping activity during the Arctic summer has increased over the past decade as sea ice has retreated. It will likely become easier and faster in the coming decades as further reductions in sea ice cover make the northern routes even more accessible. More intense Arctic shipping has significant socio-economic and political implications for global trade, northern nations, and economies strongly linked to traditional shipping corridors, while also increasing environmental risk in the Arctic. Reduced Arctic sea ice cover also allows greater access to offshore petroleum resources and ports supporting non-renewable resource extraction on land.

Melting ice sheets and glaciers in the polar regions cause sea levels to rise, affecting coastal regions and their disproportionately large populations and economies. At present, the Greenland Ice Sheet and polar glaciers are contributing more to sea level rise than the Antarctic Ice Sheet. However, changes in the Antarctic Ice Sheet have recently accelerated because of the combined effect of surface melt and basal melt from rising ocean temperatures. Even though it remains difficult to project the amount of ice loss from Antarctica after the second half of the 21st century, it is expected to contribute significantly to continued sea level rise.

The polar regions influence global climate through a number of processes. As spring snow and summer sea ice cover decrease, more heat is absorbed and the cooling effect of the polar regions on the global climate system is reduced. There is growing evidence that ongoing changes in the Arctic, primarily sea ice loss, can also influence mid-latitude weather, but more research needs to be done to identify the degree to which this is presently happening. As temperatures increase in the Arctic, permafrost soils in northern regions store less carbon. The release of carbon dioxide and methane from the land to the atmosphere further contributes to global warming.

The Southern Ocean that surrounds Antarctica is the main global region where deep waters rise to the surface, interact with the atmosphere and ice, and subsequently sink back to depth. This process stores significant amounts of heat and carbon in the deep ocean, including that produced by human activity, for decades to centuries or longer. This helps to slow the rate of global warming in the atmosphere.

[END FAQ3.1 HERE]

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Appendix 3.A: Supplementary Material

The material contained in the appendix will be presented as online supporting material to the published chapter/report.

3.A.1 Polar Regions, People and the Planet

3.A.1.1 Northern Hemispheric Climate Modes

The Northern Hemisphere atmospheric wind motion is primarily a zonal jet stream that includes multiple north-south propagating wave patterns. Recurring climate patterns can also be described using modes of atmospheric variability. The most important patterns for the Northern Hemisphere climate are centred on the North Pole, the North Atlantic, and the North Pacific.

The Arctic Oscillation (AO) or Northern Annular Mode in its positive sign has zonal symmetric flow centred on the North Pole. In its negative phase this pattern breaks down into a weaker and wavier circulation pattern. The North Atlantic Oscillation (NAO) is an Atlantic extension of the AO with a positive phase for lower pressure near Iceland.

The pattern in the North Pacific is either captured by the Pacific North-American (PNA) pattern based on geopotential height or the Pacific Decadal Oscillation (PDO) based on ocean temperature. Positive phase are associated with lower pressures in the Aleutian low region and positive temperature anomalies in the Gulf of Alaska.

Other patterns of interest is the Arctic Dipole (AD), which is the third hemispheric pattern. In contrast to the AO that is circular around a given latitudinal, the AD has flow across the central Arctic with high and low pressures on either side (Asia and North America).

The historical time series of all these patterns have inter-annual and multi-year variability that is mostly internal atmospheric stochastic variability rather than driven by external forcing such as greenhouse gas warming. The winter AO was negative up to the late 1980s (except for the early 1970s), had a large positive sign in the early 1990s, and is mostly variable since then. The PNA/PDO had a large shift in the mid-1970s and is variable and slightly positive since then. The NAO was also positive in the 1990s and variable since then. The NAO had an extreme negative winter in 2010 and an extreme positive winter in 2015. In the early 2000s a strong AD helped to reinforce summer sea ice loss (Wang et al., 2009). Since AR5 there is *medium evidence* and *medium confidence* that much variability in Northern Hemispheric atmospheric modes remains driven by internal atmospheric processes.

3.A.1.2 Arctic Amplification

The impacts of global warming are strongly manifested in the polar regions because increases in air temperature lead to reductions in snow and ice, allowing more of the sun’s energy to be absorbed by the surface, fostering more melt (Manabe and Stouffer, 1980; Overland et al., 2017) (see Chapter 3, Box 3.1). Furthermore, increased exchanges of latent heat flux from the ocean to the atmosphere has led to increased atmospheric water vapour which contributes to further warming (Serreze et al., 2012). The sea ice albedo feedback has been implicated in dramatic sea ice loss events (Perovich et al., 2008) and in the observed Arctic amplification of warming trends (Serreze et al., 2009; Screen and Simmonds, 2010) (*very high confidence*).

Modelling studies show that Arctic Amplification is related to the observed transition from perennial to seasonal sea ice (Haine and Martin, 2017), but it can still occur in the absence of the sea ice-albedo feedback (Alexeev et al., 2005) because of the contributions from other processes. There is emerging evidence of increased warm, moist air intrusions in both winter and spring (Kapsch et al., 2015; Boisvert et al., 2016; Cullather et al., 2016; Mortin et al., 2016; Graham et al., 2017). Tropical convection may play an important role by exciting these intrusion events on inter-decadal time scales (Lee et al., 2011). Intra-seasonal tropical convection variability may influence daily Arctic surface temperatures in both summer and winter (Yoo et al., 2012a; Yoo et al., 2012b; Henderson et al., 2014). The intrusion of weather events into the Arctic from the subarctic lead to increased down-welling longwave radiation from a warmer free troposphere as well as a change in optical depth from increased atmospheric moisture. A large contributor to Arctic Amplification is increased down-welling longwave radiation (Pithan and Mauritsen, 2014). It is important to clearly differentiate the contributions from local forcing (i.e., ice-albedo feedback, increased atmospheric water vapour and cloud cover) from remote forcing (i.e., changes in atmospheric circulation).

3.A.1.3 Southern Hemispheric Climate Modes

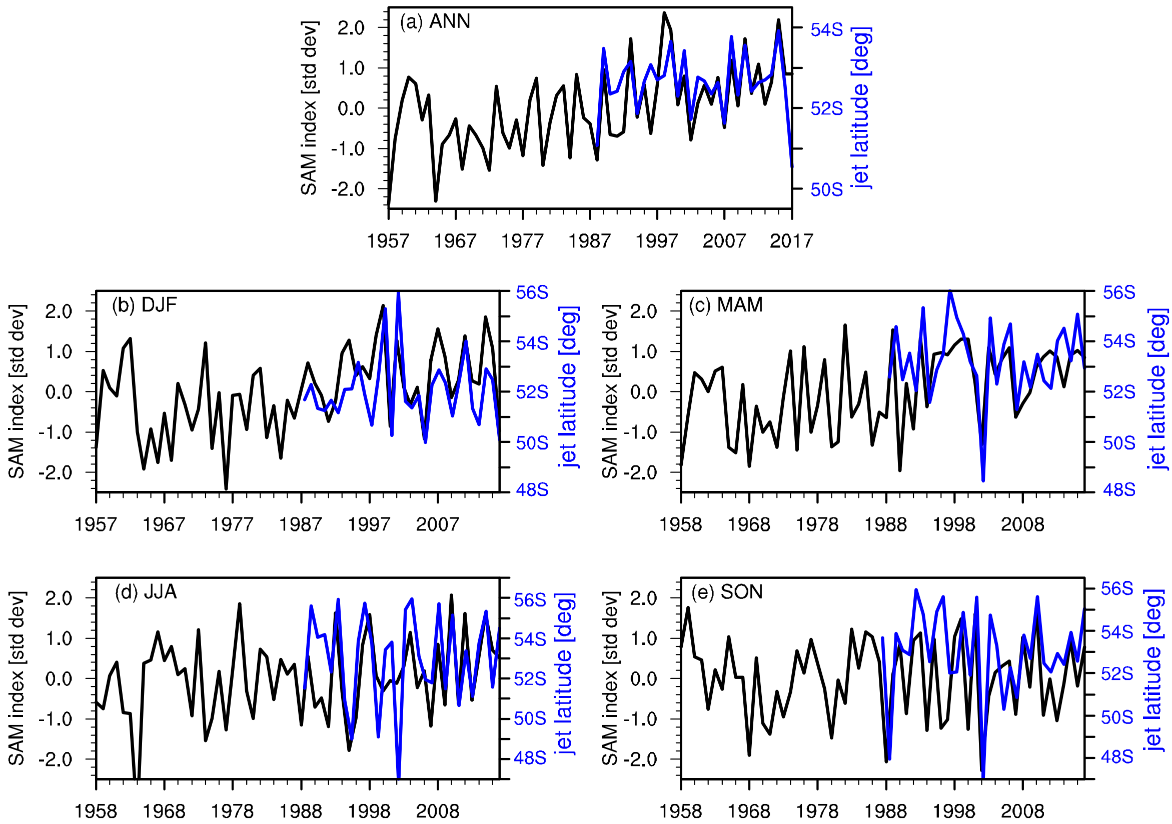
Observed changes in the Southern Hemisphere extratropical atmospheric circulation are primarily indicated by the Southern Annular Mode (SAM), the leading mode of extratropical variability in sea level pressure or geopotential heights which is related to the latitudinal position and strength of the mid-latitude eddy-driven jet. In winter and spring these winds exhibit more zonal asymmetries, expressed by the zonal wave 3 (ZW3) and Pacific South American (PSA) patterns (Irving and Simmonds, 2015). Understanding decadal variability, such as the Pacific Decadal Oscillation/Interdecadal Pacific Oscillation's (PDO/IPO) impact on these modes is hampered by the shortness of the observational record, with limited station data available poleward of 40S (Marshall, 2003).

The SAM has a strong influence on the weather and climate of SH polar regions as well as southern Australia, New Zealand, southern South America and South Africa (see review article by Thompson et al. (2011)). Numerous studies have attributed the significant poleward shift and strengthening of the SAM over the past 30-50 years to anthropogenic forcing, in particular stratospheric ozone depletion and increasing greenhouse gases (Gillett et al., 2013) (Appendix 3.A, Figure 1). Though the exact mechanisms by which these forcings impact the circulation is unclear, they both act to enhance the meridional temperature gradient which leads to a poleward shift in the SH extratropical circulation. There is *medium confidence* that ozone depletion is the dominant driver of recent austral summer changes in the Southern Hemisphere circulation during the period of maximum ozone depletion from the late 1970s to late 1990s (Arblaster et al., 2014; Waugh et al., 2015). In the years following, Waugh et al. (2015) and other studies argue for a strong impact of tropical Pacific sea surface temperatures in driving positive SAM trends (Schneider et al., 2015a; Clem et al., 2017a).

Zonal wave 3 (ZW3) describes the asymmetric part of the generally strongly zonally symmetric circulation in the SH extratropics and has been shown to impact the SH surface climate, blocking, sea-ice extent and the strength of the Amundsen Sea Low (Turner et al., 2017a; Schlosser et al., 2018). It has its strongest amplitude in SH winter and is more prominent during phases of negative SAM (Irving and Simmonds, 2015). No significant trends in the amplitude or phase of zonal wave 3 over the satellite era have been found (Turner et al., 2017c).

The Pacific South America (PSA) pattern reflects a Rossby wave train from the tropical Pacific and is the primary mechanism by which tropical Pacific SSTs, including the El Niño Southern Oscillation, impact Antarctic climate (Mo and Higgins, 1998; Irving and Simmonds, 2016). It has been shown to be closely related to the Amundsen Sea Low (Raphael et al., 2016) and to have a strong influence on temperature and precipitation variability of West Antarctica and the Antarctic Peninsula as well as sea-ice in the Amundsen, Bellingshausen and Weddell Seas (Irving and Simmonds, 2016), consistent with a deepening of the Amundsen Sea Low (Chapin III et al., 2015; Schneider et al., 2015a; Raphael et al., 2016), however there is *low confidence* in these trends and their attribution given the large internal variability in this region and shortness of the observational record.

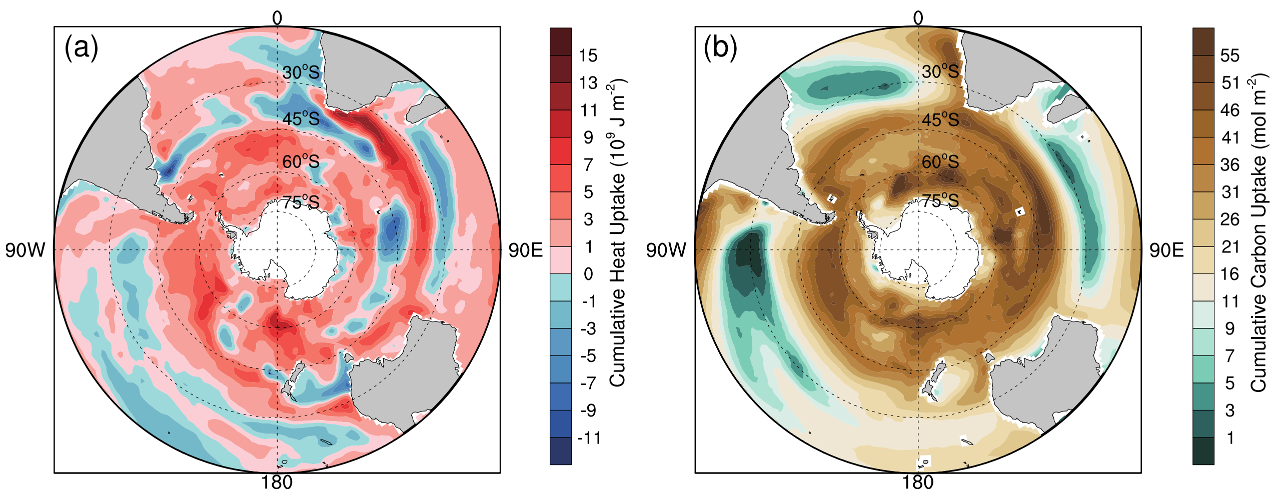
Understanding decadal variability, such as the Pacific Decadal Oscillation/Interdecadal Pacific Oscillation's (PDO/IPO) impact on these modes is hampered by the shortness of the observational record, with limited station data available poleward of 40S (Marshall, 2003) prior to the satellite era.



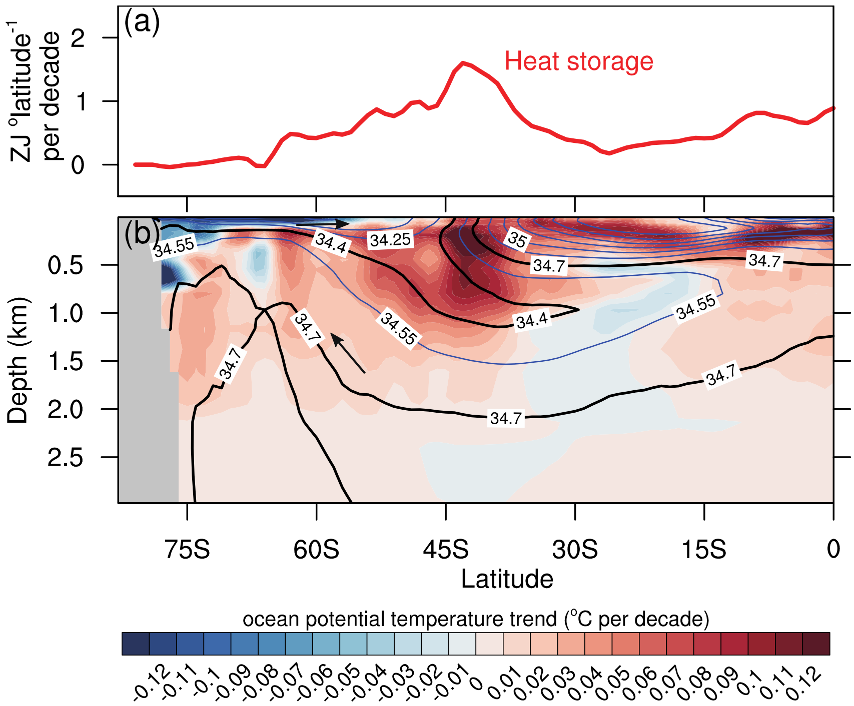
Appendix 3.A, Figure 1: SAM index (black) and mid-latitude jet positions (blue) time series for (a) annual mean and (b-e) the four seasons. The SAM index, Marshall (2003); available for download from http://www.nerc-bas.ac.uk/public/icd/gjma/newsam.1957.2007.seas.txt) is normalized by its standard deviation. The jet position is based on the maximum of CCMP satellite-based surface wind speed (Atlas et al. (2010); available for download at http://www.remss.com/measurements/ccmp.html) which starts in 1987. Adapted from Karpechko and Maycock (In press).

3.A.2 *Implications of Climate Change for Polar Oceans and Sea Ice: Feedbacks and Consequences for Ecological and Social Systems*

3.A.2.1 Heat and Carbon Uptake by the Southern Ocean

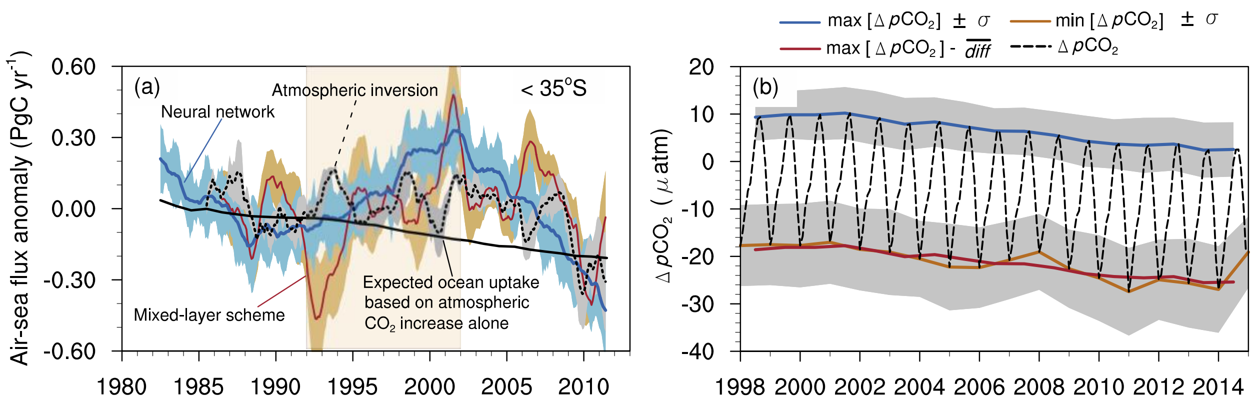


Appendix 3.A, Figure 2: CMIP5 multimodel mean changes in depth-integrated oceanic heat (a) and anthropogenic carbon (b) between 1870 (represented by mean of period 1861-80) and 1995 (represented by mean of period 1986-2005). In these models, the Southern Ocean accounts for 75 ± 22% of the total global ocean heat uptake and 43 ± 3% of anthropogenic CO_2 uptake (Frölicher et al., 2015).



Appendix 3.A, Figure 3: (a) Zonally- and depth-integrated ocean heat content trends from EN4 datasets (https://www.metoffice.gov.uk/hadobs/en4/), for period 1982-2016. (**b**) Zonal-mean ocean potential temperature trend (shading) from EN4 for 1982-2016, with climatological ocean salinity in intervals of 0.15 (contours). Arrows indicate the orientation of the residual-mean meridional overturning circulation along salinity contours 34.4 and 34.7 (heavy black lines). Updated from Armour et al. (2016).

3.A.2.2 Decadal Variability in the Southern Ocean Air-sea Flux of CO_2



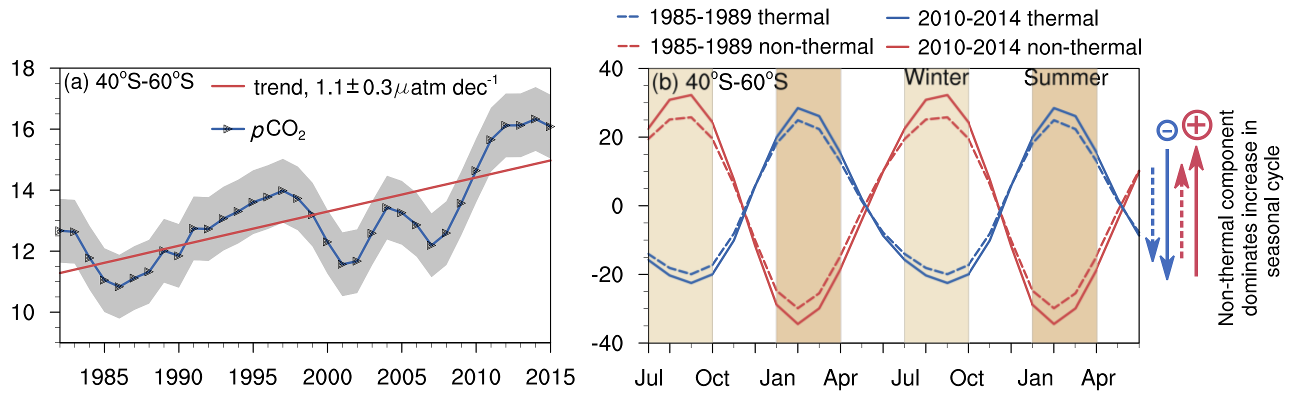
Appendix 3.A, Figure 4: (a) Decadal variability in the Southern Ocean air-sea CO_2 flux anomaly (adapted from Landschützer et al. (2015)). Curves contrast the decadal model reconstruction (1982-2012) of CO_2 air-sea flux anomalies from observations and neural network against a second empirical method (Rodenbeck et al., 2014) and a model-based steady-state linear trend of an increasing CO_2 sink. Yellow shading denotes the period of the weakening of the Southern Ocean carbon sink, separating periods of strengthening before and after. (b) The interannual variability of the seasonal cycle of ΔpCO_2 showing that the decadal trend (1998-2012) is strongly associated with trends in winter peaks of ΔpCO_2, whereas the summer minima have stronger interannual modes (adapted from Gregor et al. (2017b))

3.A.2.3 Variability and Trends in DIC Buffer Factor (γ)

The Dissolved Inorganic Carbon (DIC) buffer factor reflects the sensitivity of changing ocean pCO_2 to a changing DIC (Egleston et al., 2010). Decreasing of buffer factor or increasing Revelle Factor with rising atmospheric pCO_2 linked to anthropogenic emissions acts as a strong positive feedback on atmospheric CO_2, by reducing potential future uptake of CO_2 by the Southern Ocean (Wang et al., 2016). The Revelle Factor will grow to become one of the most important factors reducing the capacity of the Southern Ocean to take up anthropogenic CO_2 (Egleston et al., 2010) and play a positive feedback role in the carbon – climate system as well as early onset of hypercapnia or carbonate under saturation (McNeil and Sasse, 2016).

One of the important outcomes predicted by carbonate equilibrium theory for a decreasing buffering capacity is an amplified seasonal variability of pCO_2 (Egleston et al., 2010; McNeil and Sasse, 2016). A century-scale set of model runs comparing the RCP8.5 scenario with a control (constant at pre-industrial pCO_2) showed that the seasonal cycle of pCO_2 amplified by a factor of 2 – 3 mainly due to the increased sensitivity of CO_2 to summer DIC drawdown by primary productivity (Hauck and Volker, 2015). Thus in future, as buffering capacity of the ocean decreases towards the end of the century, biology will have an increased contribution to the uptake of anthropogenic carbon during the summer in the Southern Ocean (Hauck and Volker, 2015).

This has been further investigated using observation-based CO_2 products (Landschützer et al., 2018). Using the data product that spans 34 years (1982-2015) the study confirms the model predictions that there already exists an observable trend in the increase of the mean seasonal amplitude of the seasonal cycle of pCO_2 of 1.1 ± 0.3 µatm per decade in the Southern Ocean (Landschützer et al., 2018) (Appendix 3.A, Figure 5a). It also shows that this mean trend is the net effect of opposing forcing from biogeochemical (non-thermal) (2.9 ± 0.7) and thermal (-2.1 ± 0.5) (Appendix 3.A, Figure 5b). Overall, these changes to the characteristics of the seasonal cycle of biogeochemistry and CO_2 because of the trends in reduced buffering will become dominant drivers of the long-term trend of the fluxes and storage of anthropogenic CO_2 in the Southern Ocean (McNeil and Sasse, 2016).



Appendix 3.A, Figure 5: (a) The significant multi-decadal (1982:2005) trend (1.1 ± 0.3uatm/decade) in increasing amplitude of the seasonal cycle of pCO_2 in the Southern Ocean. (b)The seasonal trend signal decomposed for thermal and non-thermal drivers: non-thermal (DIC) drivers dominate the trend (b). Adapted from Landschützer et al. (2018).

3.A.2.4 Decadal Changes in Southern Ocean Carbon Storage Rates

Decadal changes in the modelled net carbon and observed anthropogenic carbon storage rates may be linked to the decadal phases of the upper-ocean overturning circulation (DeVries et al., 2017) (Appendix 3.A, Table 1). The net carbon storage is largely influenced by changes in the outgassing flux as a response to the intensification or weakening of the upwelling of circumpolar deep water. This has the potential to explain why storage increases when upper-ocean overturning weakens and outgassing is reduced (DeVries et al., 2017). In contrast, anthropogenic carbon has maximum storage during high upper-ocean overturning periods, probably due to its sensitivity to the increased rate of subduction of mode and intermediate waters (Tanhua et al., 2017). The magnitude of the carbon storage variability is therefore an indication of the sensitivity of the system to small wind-driven adjustments in the upper-ocean overturning circulation (Swart et al., 2014; Swart et al., 2015a).

Appendix 3.A, Table 1: Compares the phasing and magnitude of the decadal variability in net carbon and anthropogenic carbon storage in the Southern Ocean (DeVries et al., 2017; Tanhua et al., 2017). UOOC = Upper-Ocean Overturning Circulation; C_{ANT} = anthropogenic carbon.

Decade	DeVries et al. (2017)		Tanhua et al. (2017)	
	Net storage CO_2	Explanation	C_{ANT} Storage Rates	Explanation
1980s	High - 0.53PgCy^{-1}	Slow UOOC Outgassing reduced & storage increased	1984–1990 440 kmol yr^{-1}m^{-1}	Lower storage in mode waters
1990s	Low - 0.20PgCy^{-1}	Faster UOOC Outgassing: increased storage reduced	1984–2005 1142 kmol yr^{-1}m^{-1}	High storage in mode waters
2000s	High - 0.61PgCy^{-1}	Slow UOOC Outgassing: decreased storage increased	2005–2012 –752 kmol yr^{-1}m^{-1}	Lower storage in intermediate waters

Appendix 3.A, Table 2: Timing of the onset of monthly and annual-mean undersaturation in the Southern Ocean under different emission scenarios. The effect of the abrupt change threshold between RCP2.6 and RCP4.5/8.5 is apparent. Although all scenarios show an onset of month-long undersaturation in the 21st century, the area covered by this condition under RCP2.6 is 0.2% of that covered by RCP8.5.

Scenario	Onset of month-long undersaturation	Onset of annual undersaturation	% Impact area relative to RCP8.5
RCP8.5	2050 ± 25	+ 10–20	-
RCP4.5	2064 ± 17	+ 10–20	
RCP2.6	2033 ± 15	- None	0.2%

3.A.2.5 Climate Change Impacts on Arctic Kelp Forests

In the Arctic, biodiversity of macroalgae and biomass of kelps and associated fauna have considerably increased in the intertidal to shallow subtidal zone over the last two decades, causing changes in the food web structure and functionality. This is mostly attributed to the reduced physical impact by ice-scouring and increased light availability as a consequence of warming and concomitant fast-ice retreat (Kortsch et al., 2012; Bartsch et al., 2016; Paar et al., 2016) (*medium confidence*). Increase of summer seawater temperatures up to 10°C (IPCC 2100 scenario) will not be detrimental for Arctic kelp species. A further seawater temperature increase above 10°C which is only expected under extreme warming scenarios will definitely suppress the abundance, growth and productivity of Arctic endemic *Laminaria solidungula* and sub-Arctic *Alaria esculenta* but not of cold-temperate to Arctic *Laminaria digitata* and *Saccharina latissima* (Dieck, 1992; Gordillo et al., 2016; Roleda, 2016; Zacher et al., 2016) (*high confidence*). In total, these data support projections that kelp and macroalgal production will increase in the future Arctic (e.g., Krause-Jensen et al., 2016)). This will become more pronounced when rocky substrates hidden in current permafrost areas (Lantuit et al., 2012) will be readily colonized by kelp and other macroalgae when getting ice-free as has been verified for Antarctica (Liliana Quartino et al., 2013; Campana et al., 2017) (*high confidence*).

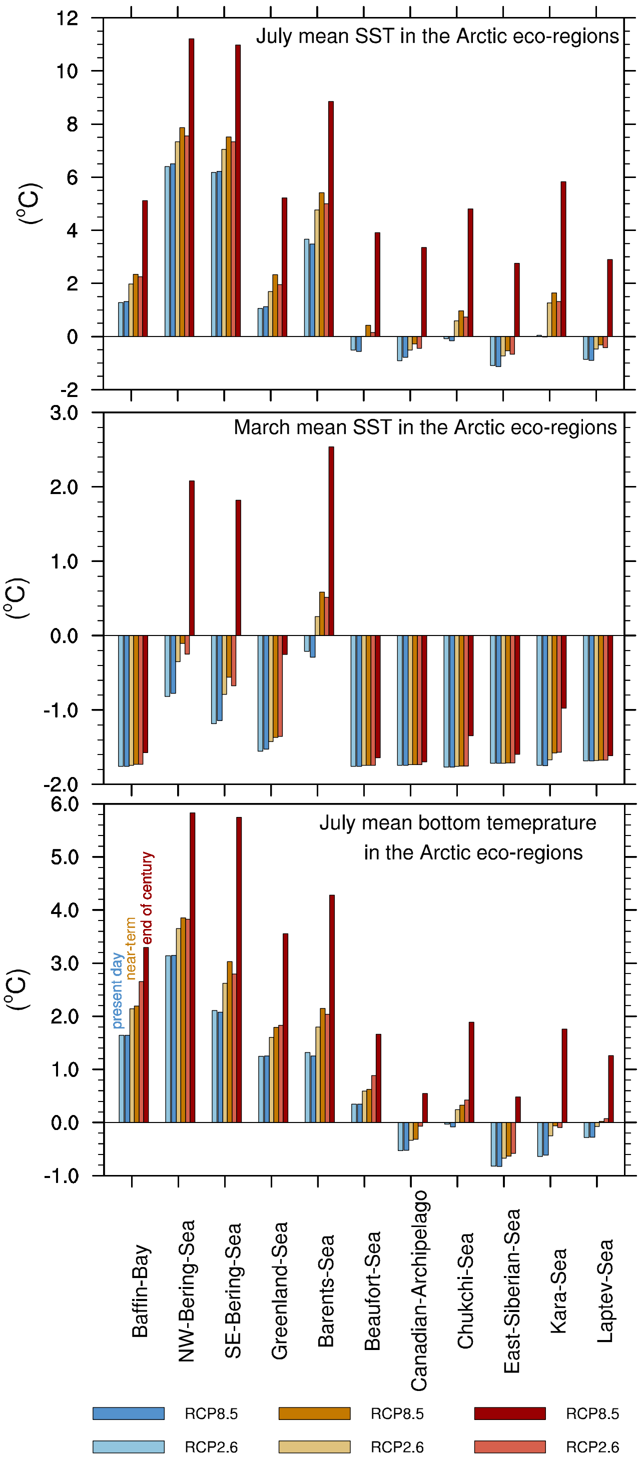
Besides the direct effects of temperature, sedimentation is a major driver in fjord systems influenced by glaciers. The reduced depth extension of several kelp species in Kongsfjorden between 1986 and 2014 was attributed to overall increased turbidity and sedimentation (Bartsch et al., 2016) (*low confidence*). Sedimentation may also inhibit the germination of Arctic kelp spores and reduce their subsequent sporophyte recruitment (*Alaria esculenta, Saccharina latissima, Laminaria digitata*). Interaction with grazing and a simulated increase in summer sea temperatures by 3°C–4°C (IPCC scenario for 2100) partially counteracts the negative impact of sedimentation in a species-specific manner (Zacher et al., 2016) (*medium confidence*). Transient sediment cover on kelp blades on the other hand provides an effective shield against harmful ultraviolet radiation (Roleda et al., 2008). Glacial melt also increases freshwater inflow into Arctic fjord systems and thereby may impose hyposaline conditions to shallow water kelps. Pre-conditioning with low salinity as a stressor results in an increased tolerance towards UV-radiation in Arctic *Alaria esculenta* thereby indicating the potential of cross-acclimation under environmental change (Springer et al., 2017).

Ocean acidification in interaction with climate warming will be most pronounced in the Arctic, where kelp and kelp like brown algae show variable species-specific responses under end of the century scenarios for CO_2 (390 and 1000 ppm) and temperature (4°C and 10°C) (Gordillo et al., 2015; Gordillo et al., 2016; Iñiguez et al., 2016). On a biochemical side, warming involves photochemistry adjustments while increased CO_2 mainly affects the carbohydrate and lipid content suggesting that ocean acidification may change metabolic pathways of carbon in kelps (Gordillo et al., 2016). Increased CO_2 also affects photosynthetic acclimation under UV radiation in Arctic *Alaria esculenta* and *Saccharina latissima* (Gordillo et al., 2015). Experimental observations support that interactions between temperature and CO_2 are low indicating a higher resilience of Arctic kelp communities to these climate drivers than their cold-temperate counterparts (Olischläger et al., 2014; Gordillo et al., 2016).

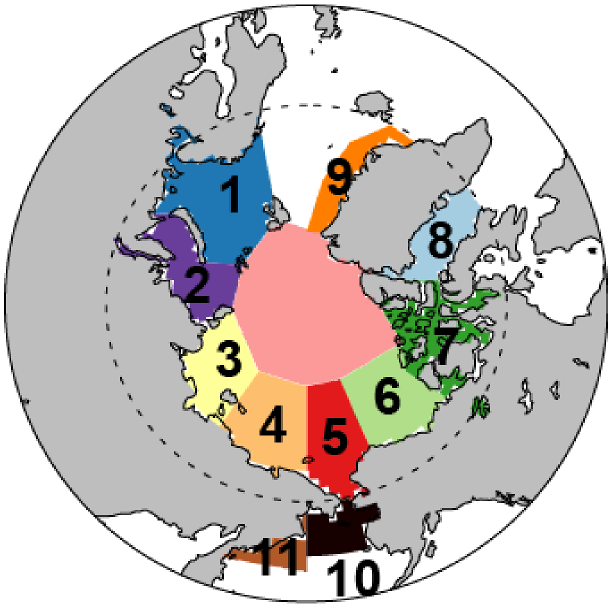
3.A.2.6 Projected Regional Changes for Sea Surface Temperature and Bottom Temperature from CMIP5 in Arctic Shelf Regions

CMIP5 ensemble projections of Arctic shelf region (bottom depths <=500 m; see Appendix 3.A, Figure 7). Bottom Temperature (BT) and Sea Surface Temperature (SST) reveal marked regional differences in the magnitude and rate of expected change in the Arctic (Appendix 3.A, Figure 6). BT and SST have direct and indirect implications to regional marine ecosystems and therefore we summarize the key findings here for reference to subsequent marine ecosystem sections:

- Projected mean SST in March exceeded 0°C by mid-century in the Barents Sea shelf region. Mean SST in March remained below 0°C in the other Arctic shelf regions in mid-century, but was projected to exceed 0°C by end of century in the eastern Bering Sea shelf (under RCP8.5) and western Bering Sea shelf (under RCP4.5 and 8.5).
- Projected mean SSTs in July at end of century were highest in the Barents Sea and Bering Sea and lowest in the East Siberian Sea (ESS) and the Canadian Archipelago (CA). With the exception of the ESS, the CA, and the Laptev Sea, mean temperature in July was projected to exceed 0°C by mid-century.
- The region experiencing the largest differences in projected mean July SST between RCP2.6 and 8.5 at end of century were in the Kara and the Chukchi Seas.
- Projected mean monthly winter BT in the East and West Greenland Seas and the Barents Seas remained above 1°C in the present, mid-century, and future. A similar pattern was projected in the Western Bering Sea. In all other regions, winter BT was projected to remain below 0°C (Kara, Laptev, East Siberian Sea, Chukchi Sea, Canadian Archipelago, or 1°C (Beaufort Sea, East Bering Sea).
- Projected mean BTs in July at end of century were highest in the Barents Sea and Bering Sea and lowest in the Kara Sea, East Siberian Sea (ESS) and the Canadian Archipelago (CA). Sub-zero temperatures in July were projected under RCP2.6 in the Kara Sea, ESS and the CA. The region experiencing the largest differences in projected mean July BT between RCP2.6 and 8.5 at end of century were in the Barents and East Bering Seas.



Appendix 3.A, Figure 6: July SST (top), March SST (middle), and July bottom temperature (bottom) for present day (2006–2015), near-term (2031–2050), and end of century (2081–2100) for RCP2.6 and 8.5.



Appendix 3.A, Figure 7: Locations of Arctic shelf regions referred to in the main text. (1) Barents Sea (2) Kara Sea (3) Laptev Sea (4) East Siberian Sea (5) Chukchi Sea (6) Beaufort Sea (7) Canadian Archipelago (8) West Greenland Sea (9) East Greenland Sea (10) East Bering Sea (11) West Bering Sea (redrawn from Carmack et al. (2015)).

3.A.3 Polar Ice Sheets and Glaciers: Changes, Consequences and Impacts

3.A.3.1 Methods of Observing Ice Sheet Changes

Since the beginning of the satellite era, frequent observations of ice sheet mass change have been made using three complementary approaches: 1) volume-change measurements from laser or radar altimetry, combined with modelled estimates of the variable density and compaction of firn and snow to calculate mass change; 2) input-output budgeting, comparing modelled surface mass balance inputs over major glacier catchments to mass outputs through glacier flux gates at or near the grounding line, using surface flow velocities estimated from radar or optical satellite images and ice thickness data; 3) changes in gravitational field over the ice sheets from satellite gravimetry. For the Greenland and Antarctic ice sheets, pre-20th century mass changes have been reconstructed using firn/ice core and geological evidence. Where possible we use this paleo evidence to support assessment of recent mass changes.

3.A.3.1.1 West Antarctica and Antarctic Peninsula

Inter-comparison between satellite methods over a common period

Comparing the three satellite methods described above for the 2003–2010 period, the estimates from altimetry, gravimetry and input-output budgeting for WAIS are –70 ± 8 Gt yr^{-1}, –101 ± 9 Gt yr^{-1} and –115 ± 43 Gt yr^{-1} (Shepherd et al., 2018) or, for a combined gravimetry-altimetry assessment, –98 ± 13 Gt yr^{-1} (Mémin et al., 2015)(*high agreement* in sign, *medium agreement* in magnitude). For the AP, the equivalent values are –10 ± 9 Gt yr^{-1}, –23 ± 5 Gt yr^{-1} and –51 ± 24 Gt yr^{-1} (Shepherd et al., 2018)(*high agreement* in sign, *medium agreement* in magnitude).

WAIS inter-comparison of satellite-derived mass changes through time

A substantial increase in WAIS mass loss reported by two multi-method studies (Bamber et al., 2018; Shepherd et al., 2018) (Appendix 3.A, Table 4) is supported by additional estimates from input-output budgeting of –34 ± 9 Gt yr^{-1} in 1979–2003, increasing to –112 ± 12 Gt yr^{-1} in 2003–2016 (Rignot et al., in review) and –214 ± 51 Gt yr^{-1} between approximately 2008 and 2015 (Gardner et al., 2018), by a satellite radar-altimetry-derived rate of –134 ± 27 Gt yr^{-1} for 2010 to 2013 (McMillan et al., 2014b), and by studies focussing on the Amundsen Sea Embayment (ASE) (below).

WAIS mass loss concentrated in the Amundsen Sea Embayment (ASE)

In the ASE, the three satellite measurement methods showed *high agreement* in both loss rates (–102 ± 10 Gt yr^{-1}) and in acceleration in loss (–15.7 ± 4.0 Gt yr^{-2}) for 2003–2011 (Sutterley et al., 2014)and, for 2003–

2013, and *high agreement* with gravimetry (–110 ± 6 Gt yr^{-1} with an acceleration of –15.1 Gt yr^{-2}) (Velicogna et al., 2014) (or a loss rate of around –120 Gt yr^{-1} given updated observations of isostatic rebound (Barletta et al., 2018)), with a statistical inversion of altimetry, gravimetry and GPS data (–102 ± 6 Gt yr^{-1}) (Martín-Español et al., 2016), and with input-output budgeting (–138 ± 42 Gt yr^{-1}) for 2008–2015 (Gardner et al., 2018).

AP inter-comparison of satellite-derived mass changes through time

On the AP, a multi-method assessment showing an increase in mass loss has from the 1990s to the last decade (Table 3.4) is supported by comparable loss estimates of –28 ± 7 Gt yr^{-1} for 2003–2013 from a statistical inversion of altimetry, gravimetry and GPS (Martín-Español et al., 2016), –31 ± 4 Gt yr^{-1} from gravimetry for 2003–2013 (with an acceleration of –3.2 ± 0.6 Gt yr^{-2}) (Velicogna et al., 2014), and from radar altimetry, –23 ± 18 Gt yr^{-1} for 2010 to 2013 (McMillan et al., 2014b) and –31 ± 29 Gt yr^{-1} for 2008–2015 (Gardner et al., 2018).

3.A.3.1.2 East Antarctic Ice Sheet

Inter-comparison between satellite methods over a common period

Altimetry, gravimetry and input-output budgeting for the 2003–2010 period for EAIS give estimates of +37 ± 18 Gt yr^{-1}, +47 ± 18 Gt yr^{-1} and –35 ± 65 Gt yr^{-1} (Shepherd et al., 2018), or, for a combined gravimetry-altimetry assessment, +51 ± 22 Gt yr^{-1} (Mémin et al., 2015) estimates that agree within uncertainties but vary in sign around zero.

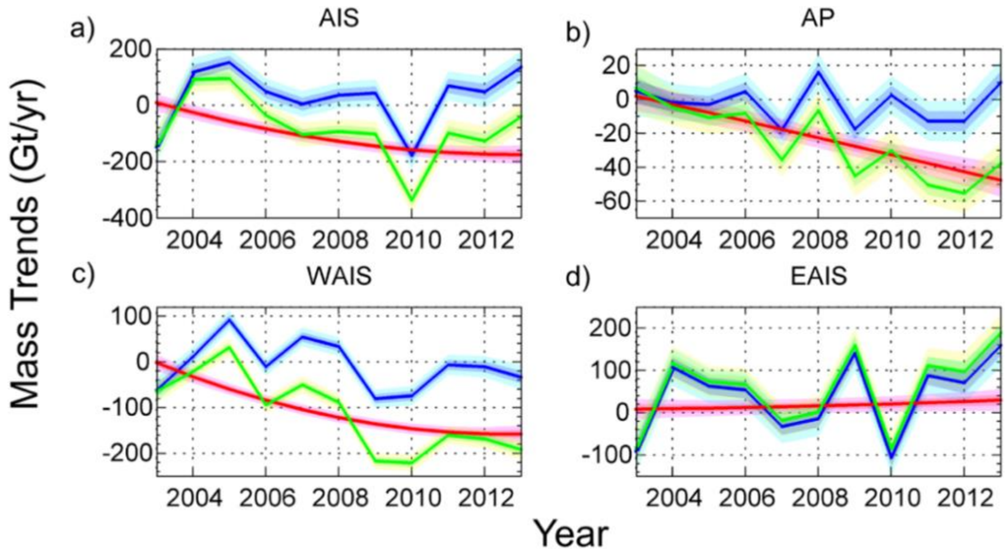
Inter-comparison of satellite-derived mass changes through time

In addition to the two multi-method satellite studies reported in Table 3.5, supporting evidence of variability but no clear multiannual trend comes from input-output budgets for EAIS ranging from –35 to +13 Gt yr^{-1} from 1979–2016(Rignot et al., in review) and +61 ± 73 Gt yr^{-1} from 2008–2015 (Gardner et al., 2018), –3 ± 36 Gt yr^{-1} from radar altimetry for 2010–2013 (McMillan et al., 2014b), and +56 ± 18 Gt yr^{-1} for 2003–2013 from a statistical inversion of altimetry, gravimetry and GPS (Zammit-Mangion et al., 2014; Martín-Español et al., 2016). One altimetry study that considered observed EAIS volume changes to be dominated by ongoing post-Holocene dynamic thickening (i.e., at the density of ice) calculated large EAIS mass gains of approximately +136 Gt yr^{-1} between 1992 and 2008 (Zwally et al., 2015), though this disagrees with other studies (Bamber et al., 2018) and was not reproduced in a sensitivity study that tested this assumption (Martin-Español et al., 2017).

3.A.3.1.3 Greenland Ice Sheet

Inter-comparison of satellite-derived mass changes through time

A multi-method satellite assessment (Table 3.6) (Bamber et al., 2018) is supported by similar results for overlapping periods from radar altimetry (–269 ± 51 Gt yr^{-1} for 2011–2016) (McMillan et al., 2016), input-output budgeting (–247 ± 28 Gt yr^{-1} for 2000–2012) (Enderlin et al., 2014) (potentially –266 Gt yr^{-1} accounting for long-term mass gains before 1990 (Colgan et al., 2015)), and gravimetry (–280 ± 58 Gt yr^{-1} for 2003–2013)(Velicogna et al., 2014).



Appendix 3.A, Figure 8: Antarctic regional mass trends for the period 2003–2013 distinguishing the SMB (blue) and ice dynamics (red) components and the total mass trend (green) for (a) Antarctic Ice Sheet, (b) Antarctic Peninsula, (c) West Antarctic Ice Sheet, and (d) East Antarctic Ice Sheet mass trends. The 1σ and 2σ confidence intervals are given by the dark and light shadings, respectively (Martín-Español et al., 2016).

Appendix 3.A, Table 4: Summary of total AIS mass balance (combined AP, WAIS and EAIS) for various periods.

Period	AIS Mass balance (Gt yr^{-1})	Uncertainty (Gt yr^{-1})	Source
2003–2010	–47	35	Mémin et al. 2014
2003–2013	–84	22	Martín-Español et al. 2015
2003–2013	–67	44	Velicogna et al. 2014
2010–2013	–160	81	McMillan et al. 2014
2008–2015	–183	94	Gardner et al. 2018
1992–2016	–93	49	Bamber et al. 2018
1992–2017	–109	56	The IMBIE team, 2018
1992–1996	–27	106	Bamber et al. 2018
1992–1997	–49	67	The IMBIE team, 2018
1997–2001	–103	157	Bamber et al. 2018
1997–2002	–38	64	The IMBIE team, 2018
2002–2006	–25	54	Bamber et al. 2018
2002–2007	–73	53	The IMBIE team, 2018
2007–2011	–117	28	Bamber et al. 2018
2007–2012	–160	50	The IMBIE team, 2018
2012–2016	–191	47	Bamber et al. 2018
2012–2017	–219	43	The IMBIE team, 2018

3.A.3.2 Projections for Polar Glaciers

Appendix 3.A, Table 5: Region-specific projected mass changes for polar glaciers at 2100 CE as a percentage change relative to modelled 2015 values. Results show multi-model means and standard deviations (SD) in response to Representative Concentration Pathway (RCP) emission scenarios. Means and SD are calculated from 6 participating glacier models forced by more than 20 General Circulation Models; results for RCP2.6 are from 46 individual glacier model simulations, while the RCP8.5 results are from 88 glacier model simulations (Hock et al., Submitted).

Region (see Figure 3.8)	RCP2.6 mean ± SD	RCP8.5 mean ± SD
Arctic Canada North (region 3)	–12 ± 7	–23 ± 14
Arctic Canada South (region 4)	–23 ± 16	–42 ± 24
Greenland (region 5)	–18 ± 10	–34 ± 15
Svalbard and Jan Mayen (region 7)	–34 ± 23	–61 ± 22
Russian Arctic (region 9)	–27 ± 21	–47 ± 27
Antarctic and Sub-Antarctic (region 19)	–12 ± 5	–25 ± 10
All polar regions excluding Antarctica	–19 ± 11	–35 ± 17
All polar regions	–18 ± 9	–32 ± 14

3.A.4 Changing Polar Seasonal Snow Cover, Permafrost and Freshwater Ice: Global and Local Impacts

[PLACEHOLDER FOR FINAL DRAFT]

3.A.5 Responding to Climate Change in Polar Systems

[PLACEHOLDER FOR FINAL DRAFT]

Acknowledgements

The authors acknowledge the following individuals for their assistance in compiling references containing Indigenous Knowledge for the Polar Regions Chapter: Claudio Aporta (Canada), David Atkinson (Canada), Todd Brinkmann (USA), Courtney Carothers (USA), Ashlee Cunsolo (Canada), Susan Crate (USA), Bathsheba Demuth (USA), Alexandra Lavrillier (France), Andrey Petrov (USA/Russia), Jon Rosales (USA), Florian Stammler (Finland), Hiroki Takakura (Japan), Wilbert van Rooij (Netherlands), Brent Wolfe (Canada), Torre Jorgenson (USA)

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