Chapter 3: Polar Regions

Coordinating Lead Authors: Michael Meredith (United Kingdom), Martin Sommerkorn (Norway/Germany)

Lead Authors: Sandra Cassotta (Denmark), Chris Derksen (Canada), Alexey Ekaykin (Russian Federation), Anne Hollowed (USA), Gary Kofinas (USA), Andrew Mackintosh (New Zealand), Jess Melbourne-Thomas (Australia), Mônica M.C. Muelbert (Brazil), Geir Ottersen (Norway), Hamish Pritchard (United Kingdom), Edward A.G. Schuur (USA)

Contributing Authors: Nerilie Abram (Australia), Julie Arblaster (Australia), Kevin Arrigo (USA), Kumiko Azetzu-Scott (Canada), David Barber (Canada), Inka Bartsch (Germany), Jeremy Bassis (USA), Dorothea Bauch (Germany), Fikret Berkes (Canada), Philip Boyd (Australia), Angelika Brandt (Germany), Lijing Cheng (China), Steven Chown (Australia), Alison Cook (United Kingdom), Jackie Dawson (Canada), Thorben Dunse (Norway/Germany), Andrea Dutton (USA), Tamsin Edwards (United Kingdom), Laura Eerkes-Medrano (Canada), Arne Eide (Norway), Howard Epstein (USA), F. Stuart Chapin III (USA), Mark Flanner (USA), Bruce Forbes (Finland), Jeremy Fyke (Canada), Andrey Glazovsky (Russian Federation), Jacqueline Grebmeier (USA), Guido Grosse (Germany), Anne Gunn (Canada), Sherilee Harper (Canada), Jan Hjort (Finland), Will Hobbs (Australia), Eric P. Hoberg (USA), Indi Hodgson-Johnston (Australia), David Holland (USA), Paul Holland (United Kingdom), Russell Hopcroft (USA), George Hunt (USA), Henry Huntington (USA), Adrian Jenkins (United Kingdom), Kit Kovacs (Norway), Gita Ljubicic (Canada), Michael Loranty (USA), Michelle Mack (USA), Andrew Meijers (United Kingdom/Australia), Benoit Meyssignac (France), Hans Meltofte (Denmark), Alexander Milner (United Kingdom), Pedro Monteiro (South Africa), Lawrence Mudryk (Canada), Mark Nuttall (Canada), Jamie Oliver (United Kingdom), James Overland (USA), Keith Reid (United Kingdom), Vladimir Romanovsky (USA/Russian Federation), Don E. Russell (Canada), Christina Schaedel (USA/Switzerland), Lars H. Smedsrud (Norway), Julienne Stroeve (Canada/USA), Alessandro Tagliabue (United Kingdom), Mary-Louise Timmermans (USA), Merritt Turetsky (Canada), Michiel van den Broeke (Netherlands), Isabella Velicogna (USA/Italy), Jemma Wadham (United Kingdom), Michelle Walvoord (USA), Gongjie Wang (China), Dee Williams (USA), Mark Wipfli (USA), Daqing Yang (Canada)

Review Editors: Oleg Anisimov (Russian Federation), Gregory Flato (Canada), Cunde Xiao (China)

Chapter Scientist: Shengping He (Norway/China), Victoria Peck (United Kingdom)

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Table of Contents

Executive Summary

This chapter assesses the state of physical, biological and social knowledge concerning the Arctic and Antarctic ocean and cryosphere, how they are affected by climate change, and how they will evolve in future. Concurrently, it assesses the local, regional and global consequences and impacts of individual and interacting polar system changes, and it assesses response options to reduce risk and build resilience in the polar regions. Key findings are:

The polar regions are losing ice, and their oceans are changing rapidly. The consequences of this polar transition extend to the whole planet, and are affecting people in multiple ways

Arctic surface air temperature has *likely*[1] increased by more than double the global average over the last two decades, with feedbacks from loss of sea ice and snow cover contributing to the amplified warming. For each of the five years since AR5 (2014-2018), Arctic annual surface air temperature exceeded that of any year since 1900. During the winters (January-March) of 2016 and 2018, surface temperatures in the central Arctic were 6°C above the 1981-2010 average, contributing to unprecedented regional sea ice absence. These trends and extremes provide *medium evidence*[2] with *high agreement* of the contemporary coupled atmosphere-cryosphere system moving well outside the 20th century envelope. {Box 3.1; 3.2.1.1}

The Arctic and Southern Oceans are continuing to remove carbon dioxide from the atmosphere and to acidify (*high confidence*). There is *medium confidence* that the amount of CO_2 drawn into the Southern Ocean from the atmosphere has experienced significant decadal variations since the 1980s. Rates of calcification (by which marine organisms form hard skeletons and shells) declined in the Southern Ocean by 3.9 ± 1.3% between 1998 and 2014. In the Arctic Ocean, the area corrosive to organisms that form shells and skeletons using the mineral aragonite expanded between the 1990s and 2010, with instances of extreme aragonite undersaturation. {3.2.1.2.4}

Both polar oceans have continued to warm in recent years, with the Southern Ocean being disproportionately and increasingly important in global ocean heat increase (*high confidence).* Over large sectors of the seasonally ice-free Arctic, summer upper mixed layer temperatures increased at around 0.5°C per decade during 1982-2017, primarily associated with increased absorbed solar radiation accompanying sea ice loss, and the inflow of ocean heat from lower latitude increased since the 2000s (*high confidence*). During 1970-2017, the Southern Ocean south of 30°S accounted for 35-43% of the global ocean heat gain in the upper 2000 m (*high confidence*), despite occupying ~25% of the global ocean area. In recent years (2005-2017), the Southern Ocean was responsible for an increased proportion of the global ocean heat increase (45-62%) (*high confidence*). {3.2.1.2.1}

Climate-induced changes in seasonal sea ice extent and thickness and ocean stratification are altering marine primary production (*high confidence*), with impacts on ecosystems (*medium confidence).*
Changes in the timing, duration and intensity of primary production have occurred in both polar oceans, with marked regional or local variability (*high confidence*). In the Antarctic, such changes have been associated with locally-rapid environmental change, including retreating glaciers and sea ice change (*medium confidence*). In the Arctic, changes in primary production have affected regional species composition, spatial distribution, and abundance of many marine species, impacting ecosystem structure (*medium confidence*). {3.2.1; 3.2.3, 3.2.4}

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[1] In this Report, the following terms have been used to indicate the assessed likelihood of an outcome or a result: Virtually certain 99–100% probability, Very likely 90–100%, Likely 66–100%, About as likely as not 33–66%, Unlikely 0–33%, Very unlikely 0–10%, and Exceptionally unlikely 0–1%. Additional terms (Extremely likely: 95–100%, More likely than not >50–100%, and Extremely unlikely 0–5%) may also be used when appropriate. Assessed likelihood is typeset in italics, e.g., *very likely* (see Section 1.9.2 and Figure 1.4 for more details). This Report also uses the term '*likely* range' to indicate that the assessed likelihood of an outcome lies within the 17-83% probability range.
[2] In this Report, the following summary terms are used to describe the available evidence: limited, medium, or robust; and for the degree of agreement: low, medium, or high. A level of confidence is expressed using five qualifiers: very low, low, medium, high, and very high, and typeset in italics, e.g., *medium confidence*. For a given evidence and agreement statement, different confidence levels can be assigned, but increasing levels of evidence and degrees of agreement are correlated with increasing confidence (see Section 1.9.2 and Figure 1.4 for more details).

In both polar regions, climate-induced changes in ocean and sea ice, together with human introduction of non-native species, have expanded the range of temperate species and contracted the range of polar fish and-ice associated species *(high confidence)*. Commercially- and ecologically-important fish stocks like Atlantic cod, haddock and mackerel have expanded their spatial distributions northwards many 100 km, and increased their abundance. In some Arctic areas, such expansions have affected the whole fish community, leading to higher competition and predation on smaller-sized fish species, while some commercial fisheries have benefited. There has been a southward shift in the distribution of Antarctic krill in the South Atlantic, the main area for the krill fishery (*medium confidence*). These changes are altering biodiversity in polar marine ecosystems (*medium confidence*) {3.2.3; Box 3.4}.

It is *very likely* that Arctic sea ice extent continues to decline in all months of the year; the strongest reductions in September (-12.8 ± 2.3% per decade; 1979-2018) are *likely* unprecedented in at least 1000 years. It is *virtually certain* that Arctic sea ice has thinned, concurrent with a shift to younger ice: since 1979, the areal proportion of thick ice at least 5 years old has declined by approximately 90%. It is *very likely* that approximately half the observed sea ice loss is attributable to increased atmospheric greenhouse gas concentrations. Changes in Arctic sea ice have potential to influence midlatitude weather on timescales of weeks to months (*medium confidence*). {3.2.1.1; Box 3.2}

It is *very likely* that Antarctic sea ice cover exhibits no significant trend over the period of satellite observations (1979 to 2018). While the drivers of historical decadal variability are known with *medium confidence,* there is currently *limited evidence* and *low agreement* concerning causes of the strong recent decrease (2016-2018), and *low confidence* in the ability of current-generation climate models to reproduce and explain the observations. {3.2.1.1}

Shipping activity during the Arctic summer increased over the past two decades in regions for which there is information, concurrent with reductions in sea ice extent (*high confidence*). Transit times across the Northern Sea Route have shortened due to lighter ice conditions, and while long-term, pan-Arctic datasets are incomplete, the distance travelled by ships in Arctic Canada nearly tripled during 1990-2015 (*high confidence*). Greater levels of Arctic ship-based transportation and tourism have socio-economic and political implications for global trade, northern nations, and economies linked to traditional shipping corridors; they will also exacerbate region-specific risks for marine ecosystems and coastal communities if further action to develop and adequately implement regulations does not keep pace with increased shipping (*high confidence*). {3.2.1.1; 3.2.4.2; 3.2.4.3; 3.4.3.3.2; 3.5.2.7}

Permafrost temperatures have increased to record high levels (*very high confidence*), but there is *medium evidence* and *low agreement* that this warming is currently causing northern permafrost regions to release additional methane and carbon dioxide. During 2007-2016, continuous-zone permafrost temperatures in the Arctic and Antarctic increased by 0.39 ± 0.15°C and 0.37 ± 0.10°C respectively. Arctic and boreal permafrost region soils contain 1440-1600 Gt organic carbon (*medium confidence*). Changes in permafrost influence global climate through emissions of carbon dioxide and methane released from the microbial breakdown of organic carbon, or the release of trapped methane. {3.4.1; 3.4.3}

Climate-related changes to Arctic hydrology, wildfire and abrupt thaw are occurring (*high confidence*), with impacts on vegetation and water and food security. Snow and lake ice cover has declined, with June snow extent decreasing 13.4 ± 5.4% per decade (1967-2018) (*high confidence*). Runoff into the Arctic Ocean increased for Eurasian and North American rivers by 3.3 ± 1.6% and 2.0 ± 1.8% respectively (1976-2018; *medium confidence*). Area burned and frequency of fires (including extreme fires) are unprecedented over the last 10,000 years (*high confidence*). There has been an overall greening of the tundra biome, but also browning in some regions of tundra and boreal forest, and also changes in the abundance and distribution of animals including reindeer and salmon (*high confidence*). Together, these impact access to (and food availability within) herding, hunting, fishing, forage and gathering areas, affecting the livelihood, health and cultural identity of residents including Indigenous peoples (*high confidence*). {3.4.1; 3.4.3; 3.5.2}

Limited knowledge, financial resources, human capital and organisational capacity are constraining adaptation in many human sectors in the Arctic (*high confidence*). Harvesters of renewable 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 in some cases planning for relocation (*high confidence*). In spite of these adaptations, many groups are making decisions without adequate knowledge to forecast near- and long-term conditions, and without the funding, skills and institutional support to engage fully in planning processes (*high confidence*). {3.5.2, 3.5.4, Cross-Chapter Box 9}

It is *extremely likely* that the rapid ice loss from the Greenland and Antarctic ice sheets during the early 21st century has increased into the near present-day, adding to the ice sheet contribution to global sea level rise. From Greenland, the 2012-2016 ice losses (-247 ± 15 Gt yr^{-1}) were similar to those from 2002-2011 (-263 ± 21 Gt yr^{-1}) and *extremely likely* greater than from 1992-2001 (-8 ± 82 Gt yr^{-1}). Summer melting of the Greenland Ice Sheet has increased since the 1990s (*very high confidence*) to a level unprecedented over at least the last 350 years, and two-to-fivefold the pre-industrial level (*medium confidence*). From Antarctica, the 2012-2016 losses (-199 ± 26 Gt yr^{-1}) were *extremely likely* greater than those from 2002-2011 (-82 ± 27 Gt yr^{-1}) and *likely* greater than from 1992-2001 (-51 ± 73 Gt yr^{-1}). Antarctic ice loss is dominated by acceleration, retreat and rapid thinning of major West Antarctic Ice Sheet outlet glaciers (*very high confidence*), driven by melting of ice shelves by warm ocean waters (*high confidence*). The combined sea level rise contribution from both ice sheets for 2012-2016 was 1.2 ± 0.1 mm yr^{-1}, a 29% increase on the 2002-2011 contribution and a ~700% increase on the 1992-2001 period. {3.3.1}

Mass loss from Arctic glaciers (-212 ± 29 Gt yr^{-1}) during 2006-2015 contributed to sea level rise at a similar rate (0.6 ± 0.1 mm yr^{-1}) to the Greenland Ice Sheet (*high confidence*).
Over the same period in Antarctic and subantarctic regions, glaciers separate from the ice sheets changed mass by -11 ± 108 Gt yr^{-1} (*low confidence*). {2.2.3, 3.3.2}

There is *limited evidence* and *high agreement* that recent Antarctic Ice Sheet mass losses could be irreversible over decades to millennia. Rapid mass loss due to glacier flow acceleration in the Amundsen Sea Embayment of West Antarctica and in Wilkes Land, East Antarctica, may indicate the beginning of Marine Ice Sheet Instability, but observational data are not yet sufficient to determine whether these changes mark the beginning of irreversible retreat. {3.3.1; Cross-Chapter Box 8 in Chapter 3; 4.2.3.1.2}

The polar regions will be profoundly different in future compared with today, and the degree and nature of that difference will depend strongly on the rate and magnitude of global climatic change[3]. This will challenge adaptation responses regionally and worldwide.

It is *very likely* that projected Arctic warming will result in continued loss of sea ice and snow on land, and reductions in the mass of glaciers. Important differences in the trajectories of loss emerge from 2050 onwards, depending on mitigation measures taken (*high confidence*). For stabilised global warming of 1.5°C, an approximately 1% chance of a given September being sea ice free at the end of century is projected; for stabilised warming at a 2°C increase, this rises to 10-35% (*high confidence*). The potential for reduced (further 5-10%) but stabilised Arctic autumn and spring snow extent by mid-century for RCP2.6 contrasts with continued loss under RCP8.5 (a further 15-25% reduction to end of century) (*high confidence*). Projected mass reductions for polar glaciers between 2015 and 2100 range from $16 \pm 7\%$ for RCP2.6 to $33 \pm 11\%$ for RCP8.5 (*medium confidence*). {3.2.2; 3.3.2; 3.4.2, Cross-Chapter Box 6 in Chapter 2}

Both polar oceans will be increasingly affected by CO_2 uptake, causing conditions corrosive for calcium carbonate shell-producing organisms (*high confidence*), with associated impacts on marine organisms and ecosystems (*medium confidence*). It is *very likely* that both the Southern Ocean and the Arctic Ocean will experience year-round conditions of surface water undersaturation for mineral forms of calcium carbonate by 2100 under RCP8.5; under RCP2.6 the extent of undersaturated waters are reduced markedly. Imperfect representation of local processes and sea-ice interaction in global climate models limit the ability to project the response of specific polar areas and the precise timing of undersaturation at seasonal scales. Differences in sensitivity and the scope for adaptation to projected levels of ocean acidification exist across a broad range of marine species groups. {3.2.1; 3.2.2.3; 3.2.3}

[3] Projections for ice sheets and glaciers in the polar regions are summarized in Chapters 4 and 2, respectively.

Future climate-induced changes in the polar oceans, sea ice, snow and permafrost will drive habitat and biome shifts, with associated changes in the ranges and abundance of ecologically-important species (*medium confidence*). Projected shifts will include further habitat contraction and changes in abundance for polar species, including marine mammals, birds, fish, and Antarctic krill (*medium confidence*). Projected range expansion of subarctic marine species will increase pressure for high-Arctic species (*medium confidence*), with regionally-variable impacts. Continued loss of Arctic multi-year sea ice will affect ice-related and pelagic primary production (*high confidence*), with impacts for whole ice-associated, seafloor and open ocean ecosystems. On Arctic land, projections indicate a loss of globally-unique biodiversity as some high-Arctic species will be outcompeted by more temperate species and very limited refugia exist (*medium confidence).* Woody shrubs and trees are projected to expand, covering 24-52% of the current tundra region by 2050. {3.2.2.1; 3.2.3; 3.2.3.1; Box 3.4; 3.4.2; 3.4.3}

The projected effects of climate-induced stressors on polar marine ecosystems present risks for commercial and subsistence fisheries with implications for regional economies, cultures and the global supply of fish, shellfish, and Antarctic krill (*high confidence*). Future impacts for linked human systems depend on the level of mitigation and especially the responsiveness of precautionary management approaches (*medium confidence*). Polar regions support several of the world's largest commercial fisheries. Specific impacts on the stocks and economic value in both regions will depend on future climate change and on the strategies employed to manage the effects on stocks and ecosystems (*medium confidence*). Under high emission scenarios current management strategies of some high-value stocks may not sustain current catch levels in the future (*low confidence*); this exemplifies the limits to the ability of existing natural resource management frameworks to address ecosystem change. Adaptive management that combines annual measures and within-season provisions informed by assessments of future ecosystem trends reduces the risks of negative climate change impacts on polar fisheries (*medium confidenc*e). {3.2.4; 3.5.2; 3.5.4}

Widespread disappearance of Arctic near-surface permafrost is projected to occur this century as a result of warming (*high confidence*), with important consequences for global climate. By 2100, near-surface permafrost area will decrease by 2-66% for RCP2.6 and 30–99% for RCP8.5. This could release 10s to 100s of Gt C as carbon dioxide and methane to the atmosphere for RCP8.5, with the potential to accelerate climate change (*medium confidence*). Methane will contribute a small proportion of these additional carbon emissions, on the order of 0.01-0.06 Gt CH_4 yr^{-1}, but could contribute 40-70% of the total permafrost-affected radiative forcing because of its higher warming potential. There is *medium evidence* but with *low agreement* whether the level and timing of increased plant growth and replenishment of soil will compensate these permafrost carbon losses. {3.4.2; 3.4.3}

Projected permafrost thaw and decrease in snow will affect Arctic hydrology and wildfire, with impacts on vegetation and human infrastructure (*medium confidence).* About 20% of Arctic land permafrost is vulnerable to abrupt permafrost thaw and ground subsidence, which is expected to increase small lake area by over 50% by 2100 for RCP8.5 (*medium confidence*). Even as the overall regional water cycle intensifies, including increased precipitation, evapotranspiration, and river discharge to the Arctic Ocean, decreases in snow and permafrost may lead to soil drying (*medium confidence*). Fire is projected to increase for the rest of this century across most tundra and boreal regions, while interactions between climate and shifting vegetation will influence future fire intensity and frequency (*medium confidence*). By 2050, 70% of Arctic infrastructure is located in regions at risk from permafrost thaw and subsidence; adaptation measures taken in advance could reduce costs arising from thaw and other climate-change related impacts such as increased flooding, precipitation, and freeze-thaw events by half (*medium confidence*). {3.4.1; 3.4.2; 3.4.3; 3.5.2}.

Response options exist that can ameliorate the impacts of polar change, build resilience and allow time for effective mitigation measures. Institutional barriers presently limit their efficacy.

Responding to climate change in polar regions will be more effective if attention to reducing immediate risks (short-term adaptation) is concurrent with long-term planning that builds resilience to address expected and unexpected impacts (*high confidence*). Emphasis on short-term adaptation to specific problems will ultimately not succeed in reducing the risks and vulnerabilities to society given the scale, complexity and uncertainty of climate change. Moving toward a dual focus of short- and long-term adaptation involves knowledge co-production, linking knowledge with decision-making and implementing ecosystem-based stewardship, which involves the transformation of many existing institutions (*high confidence*). {3.5.4}

Innovative tools and practices in polar resource management and planning show strong potential in improving society's capacity to respond to climate change (*high confidence*). Networks of protected areas, participatory scenario analysis, decision-support systems, community-based ecological monitoring that draws on local and indigenous knowledge, and self-assessments of community resilience contribute to strategic plans for sustaining biodiversity and limit risk to human livelihoods and wellbeing. Such practices are most effective when linked closely to the policy process. Experimenting, assessing, and continually refining practices while strengthening the links with decision making has the potential to ready society for the expected and unexpected impacts of climate change (*high confidence*). {3.5.1, 3.5.2, 3.5.4}

Institutional arrangements that provide for strong multiscale linkages with Arctic local communities can benefit from including indigenous knowledge and local knowledge in the formulation of adaptation strategies (*high confidence*). The tightly-coupled relationship of northern local communities and their environment provide an opportunity to better understand climate change and its effects, support adaptation and limit unintended consequences. Enabling conditions for the involvement of local communities in climate adaptation planning include investments in human capital, engagement processes for knowledge co-production, and systems of adaptive governance. {3.5.3}

The capacity of governance systems in polar regions to respond to climate change has strengthened recently, but the development of these systems is not sufficiently rapid or robust to address the challenges and risks to societies posed by projected changes (*high confidence*). Human responses to climate change in the polar regions occur in a fragmented governance landscape. Climate change, new polar interests from outside the regions, and an increasingly-active role played by informal organisations are compelling stronger coordination and integration between different levels and sectors of governance. The governance landscape is currently not sufficiently equipped to address cascading risks and uncertainty in an integrated and precautionary way within existing legal and policy frameworks (*high confidence*). {3.5.3, 3.5.4}

3.1 Introduction: Polar Regions, People and the Planet

This chapter provides an integrated assessment of climate change across the physical, biological and human dimensions of the polar regions, based on emerging understanding that assessing these dimensions in isolation is not sufficient or forward-looking. This offers the opportunity, for the first time in a global report, to trace cause and consequence of climate change from polar ocean and cryosphere systems to biological and social impacts, and relate them to responses to reduce risks and enhance adaptation options and resilience. To achieve this, the chapter draws on the body of literature and assessments pertaining to climate-induced dynamics and functioning of the polar regions published since the IPCC's Fifth Assessment Report (AR5), which has expanded considerably motivated in large part by growing appreciation of the importance of these regions to planetary systems and to the lives and livelihoods of people across the globe.

As integral parts of the Earth System, the polar regions interact with the rest of the world through shared ocean, atmosphere, ecological and social systems; notably, they are key components of the global climate system. This chapter therefore takes a systems approach that emphasises the interactions of cryosphere and ocean changes and their diverse consequences and impacts to assess key issues of climatic change for the polar regions, the planet and its people (Figure 3.1).

The spatial footprints of the polar regions (Figure 3.2) 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, 69% of the world's glacier area including both of the world's ice sheets, almost all of the world's sea ice, and land areas with the most persistent winter snow cover.

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 interactions of cryosphere and ocean systems and their global linkages. The different physical settings have also led to the evolution of unique marine and terrestrial biology in each polar region and shape effects, impacts and adaptation of polar ecosystems.

It is important to recognise the existence of multiple and diverse perspectives of the polar regions, many of them overlapping. These multiple perspectives encompass the polar regions as a source of resources, a key part of the global climate system, a place for preserving intact ecosystems, a place for international cooperation and, importantly, a homeland. While many of these perspectives are equally relevant for both polar regions, only the Arctic has a population for whom the region is a permanent home: approximately four million people reside there, of whom 10% are indigenous. By contrast, the Antarctic population changes seasonally between approximately 1100 and 4400, based predominantly at research stations. When assessing knowledge relating to climate change in the context of adaptation options, limits and enhancing resilience (Cross-Chapter Box 2 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. Indigenous knowledge and local knowledge are different and unique sources of knowledge that are increasingly recognised to contribute to observing, understanding, and responding to climate-induced changes (Cross-Chapter Box 4 in Chapter 1). Considering indigenous knowledge and local knowledge facilitates cooperation in the development, identification, and decision-making processes for responding to climate change in communities across the Arctic, and better understanding of the challenges facing Indigenous peoples. This chapter incorporates published indigenous knowledge and local knowledge for assessing climate change impacts and responses.

Introduction:

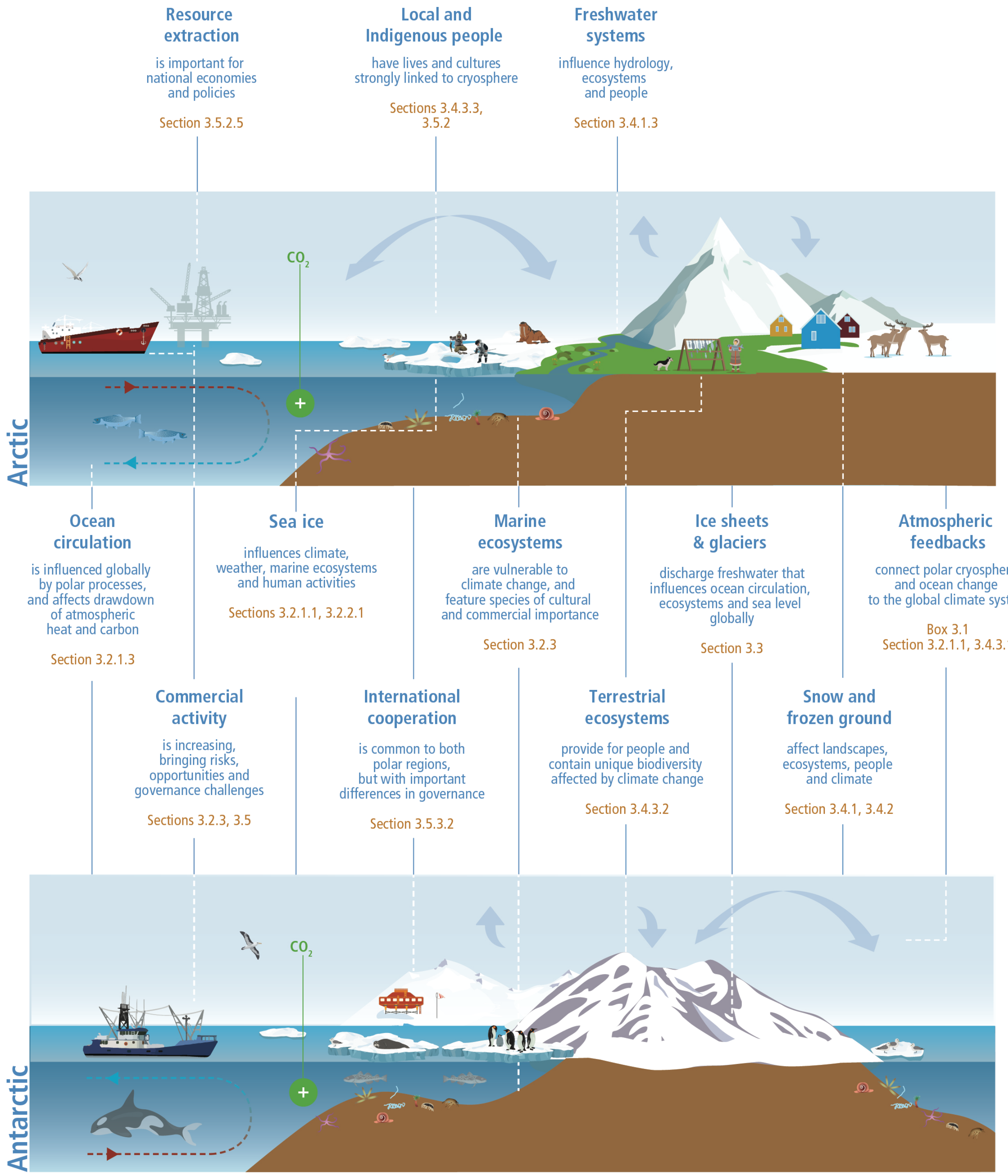


Figure 3.1: Schematic of some of the key features and mechanisms assessed in this Chapter, and by which the cryosphere and ocean in the polar regions influence climate, ecological and social systems in the regions and across the globe. Specific elements are labelled, and section numbers given for where detailed assessment information can be found.

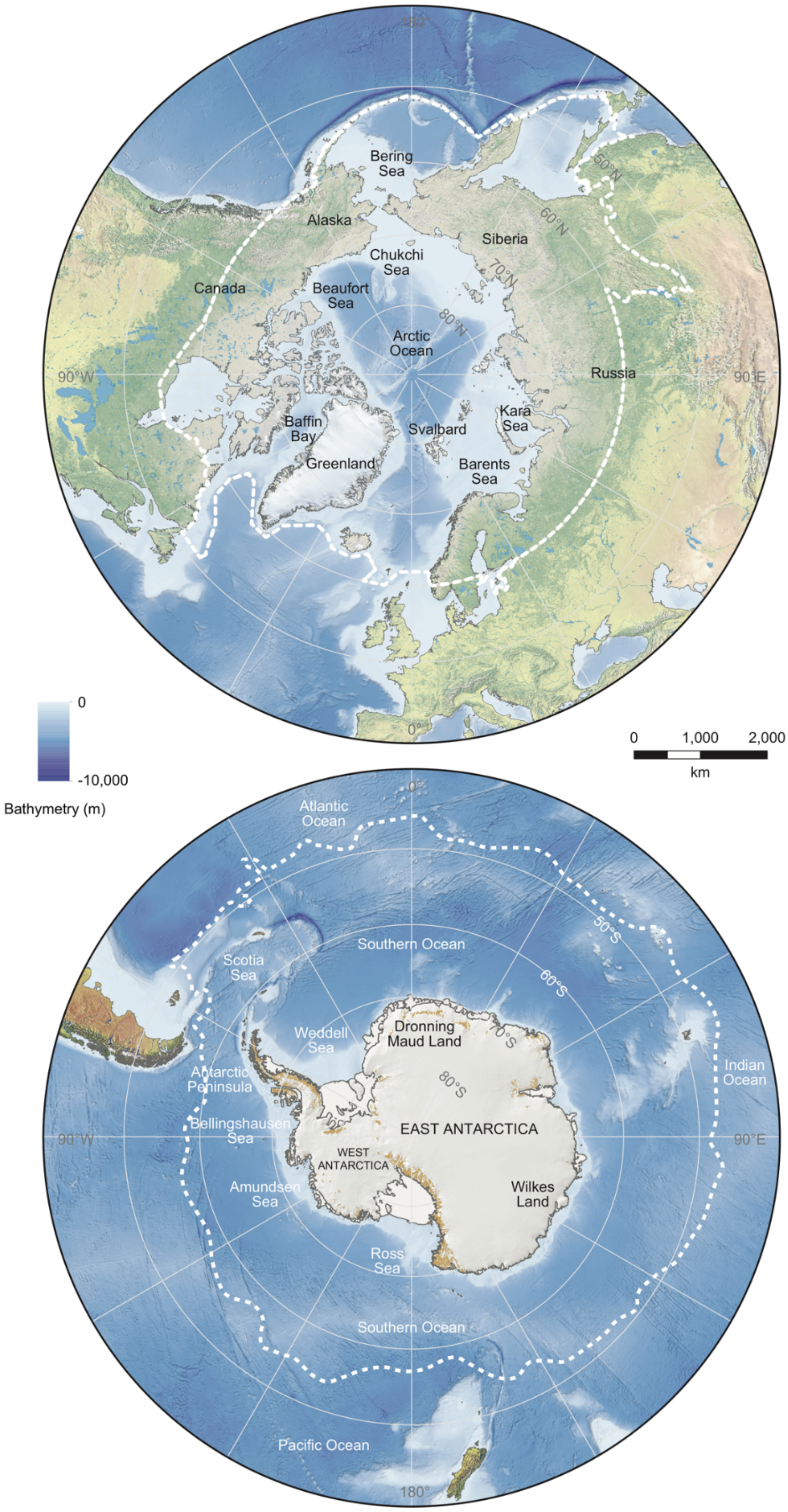


Figure 3.2: The Arctic (top) and Antarctic (bottom) polar regions. Various place names referred to in the text are marked. Dashed lines denote approximate boundaries for the polar regions; as their spatial footprint varies in relation to particular cryosphere and ocean elements or scientific disciplines, this chapter adopts a purposefully flexible approach to their delineation. The southern polar region encompasses the flow of the Antarctic Circumpolar Current at least as far north as the Subantarctic Front and fully encompasses the Convention for the Conservation of Antarctic Marine Living Resources Statistical Areas (CCAMLR, 2017c), the Antarctic continent and Antarctic and subantarctic islands, whilst the marine Arctic includes the areas of the Arctic Large Marine Ecosystems (PAME, 2013). The terrestrial Arctic comprises the areas of the northern continuous and discontinuous permafrost zone, the Arctic biome inclusive of glacial ice, and the parts of the boreal biome that are characterised by cryosphere elements, such as permafrost and persistent winter season snow cover.

[START BOX 3.1 HERE]

Box 3.1: Polar Region Climate Trends

Over the last two decades, it is *likely* that Arctic surface air temperature has increased at more than double the global average (Notz and Stroeve, 2016; Richter-Menge et al., 2017). Attribution studies show the important role of anthropogenic increases in greenhouse gases in driving observed Arctic surface temperature increases (Fyfe et al., 2013; Najafi et al., 2015), so there is *high confidence* in projections of further Arctic warming (Overland et al., 2018a). Mechanisms for Arctic amplification are still debated, but include: reduced summer albedo due to sea ice and snow cover loss, the increase of total water vapour content in the Arctic atmosphere, changes in total cloudiness in summer, additional heat generated by newly formed sea ice across more extensive open water areas in the autumn, northward transport of heat and moisture and 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; Stuecker et al., 2018) (SM3.1.1).

A number of recent events in the Arctic indicate new extremes in the Arctic climate system. Annual Arctic surface temperature for each of the past five years since AR5 (2014-2018; relative to a 1980-2010 base line) exceeded that of any year since 1900 (Overland et al., 2018b). Winter (January-March) near-surface temperature anomalies of +6°C (relative to 1981-2010) were recorded in the central Arctic during both 2016 and 2018, nearly double the previous record anomalies (Overland and Wang, 2018a). These events were caused by a split of the tropospheric polar vortex into two cells, which facilitated the intrusion of subarctic storms (Overland and Wang, 2016). The resulting advection of warm air and moisture from the Pacific and Atlantic Oceans into the central Arctic increased downward longwave radiation, delayed sea ice freeze-up, and contributed to 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). In addition to dramatic Arctic summer sea ice loss over the past 15 years, all Arctic winter sea ice maxima of the last 4 years were at record low levels relative to 1979-2014 (Overland, 2018). Multi-year, large magnitude extreme positive Arctic temperatures and sea ice minimums (Section 3.2.1.1) since AR5 provide *high agreement* and *medium evidence* of contemporary conditions well outside the envelope of previous experience (1900-2017) (AMAP, 2017d; Walsh et al., 2017).

In contrast to the Arctic, the Antarctic continent has seen less uniform temperature changes over the past 30-50 years, with warming over parts of West Antarctica and no significant overall change over East Antarctica (Nicolas and Bromwich, 2014; Jones et al., 2016; Turner et al., 2016), though there is *low confidence* in these changes given the sparse *in situ* records and large interannual to interdecadal variability. This weaker amplified warming compared to the Arctic is due to deep ocean mixing and ocean heat uptake over the Southern Ocean (Collins et al., 2013). The Southern Annular Mode (SAM), 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 (SM3.1.3). Over recent decades the SAM has exhibited a positive trend during austral summer, indicating a strengthening of the surface westerly winds around Antarctica. This extended positive phase of the SAM is unprecedented in at least 600 years, according to paleoclimate reconstructions (Abram et al., 2014; Dätwyler et al., 2017) and is associated with cooler conditions over the continent.

Consistent with AR5, it is *likely* that Antarctic ozone depletion has been the dominant driver of the positive trend in the SAM during austral summer from the late 1970s to the late 1990s (Schneider et al., 2015; Waugh et al., 2015; Karpechko et al., 2018), the period during which ozone depletion was increasing. There is *high confidence* through a growing body of literature that variability of tropical sea surface temperatures can influence Antarctic temperature changes (Li et al., 2014; Turner et al., 2016; Clem et al., 2017; Smith and Polvani, 2017) and the Southern Hemisphere mid-latitude circulation (Li et al., 2015a; Raphael et al., 2016; Turney et al., 2017; Evtushevsky et al., 2018; Yuan et al., 2018). New research suggests a stronger role of tropical sea surface temperatures in driving changes in the SAM since 2000 (Schneider et al., 2015; Clem et al., 2017).

[END BOX 3.1 HERE]

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

3.2.1 Observed Changes in Sea Ice and Ocean

3.2.1.1 Sea Ice

Sea ice reflects a high proportion of incoming solar radiation back to space, provides thermal insulation between the ocean and atmosphere, influences thermohaline circulation, and provides habitat for ice-associated species. Sea ice characteristics differ between the Arctic and Antarctic. Expansion of winter sea ice in the Arctic is limited by land, and ice circulates within the central Arctic basin, some of which survives the summer melt season to form multi-year ice. Arctic sea ice variability and impacts on communities includes indigenous knowledge and local knowledge from across the circumpolar Arctic (Cross-Chapter Box 3 in Chapter 1). The Antarctic continent is surrounded by sea ice which interacts with adjacent ice shelves; winter season expansion is limited by the influence of the Antarctic Circumpolar Current.

3.2.1.1.1 Extent and concentration
The pan-Arctic loss of sea ice cover is a prominent indicator of climate change. It is *very likely* that sea ice extent (the total area of the Arctic with at least 15% sea ice concentration) has declined since 1979 in each month of the year (Barber et al., 2017; Comiso et al., 2017b; Stroeve and Notz, 2018) (Figure 3.3). Changes are largest in summer and smallest in winter, with the strongest trends in September (1979-2018; summer month with the lowest sea ice cover) of –83,000 km^2 yr^{-1} (-12.8% per decade +/- 2.3% relative to 1981-2010 mean), and –41,000 km^2 yr^{-1} (–2.7% per decade +/-0.5% relative to 1981-2010 mean) for March (1979-2019; winter 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 et al., 2018). 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). Summer Arctic sea ice loss since 1979 is unprecedented in 150 years based on historical reconstructions (Walsh et al., 2017) and more than 1,000 years based on paleoclimate evidence (Polyak et al., 2010; Kinnard et al., 2011; Halfar et al., 2013) (*medium confidence*).

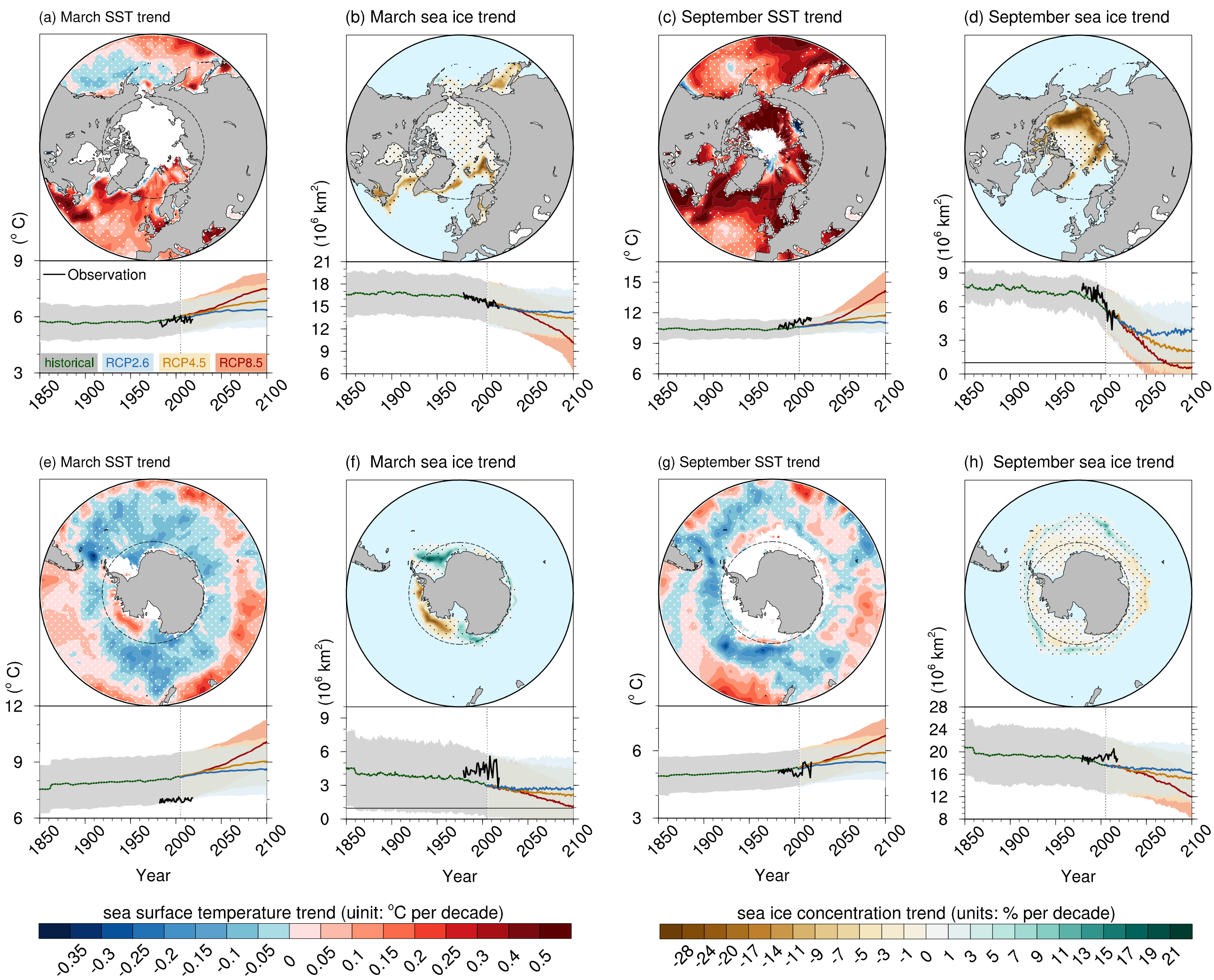


Figure 3.3: Maps of linear trends (in °C per decade) of Arctic (a, c) and Antarctic (e, g) sea surface temperature (SST) for 1982−2017 in March (a, e) and September (c, g). (b, d, f, h) same as (a, c, e, g), but for the linear trends of sea ice concentration (in % per decade). Stippled regions indicate the trends that are statistically insignificant. Dashed circles indicate the Arctic/Antarctic Circle. Beneath each map of linear trend shows the time series of SST (area-averaged north of 40°N/south of 40°S) or sea ice extent in the northern/southern hemisphere. Black, green, blue, orange, and red curves indicate observations, CMIP5 historical simulation, RCP2.6, RCP4.5, and RCP8.5 projections respectively; shading indicates +/- standard deviation of multi-models. SST trend was calculated from Hadley Centre Sea Ice and Sea Surface Temperature data set (Version 1, HadISST1; Rayner, 2003). Sea ice concentration trend was calculated from the NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, Version 3 (https://nsidc.org/data/g02202). The time series of observed SST are averages of HadISST1 and NOAA Optimum Interpolation Sea Surface Temperature dataset (version 2; Reynolds et al., 2002). The time series of observed sea ice extent are the averages of HadISST, the NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, and the Global sea ice concentration reprocessing dataset from EUMETSAT (http://osisaf.met.no/p/ice/ice_conc_reprocessed.html).

Approximately half of the observed Arctic summer sea ice loss is driven by increased concentrations of atmospheric greenhouse gases, with the remainder attributed to internal climate variability (Kay et al., 2011; Notz and Marotzke, 2012) (*medium confidence*). The sea ice albedo feedback (increased air temperature reduces sea ice cover, allowing more energy to be absorbed at the surface, fostering more melt) is a key driver of sea ice loss (Perovich and Polashenski, 2012; Stroeve et al., 2012b; Serreze et al., 2016) and is exacerbated by the transition from perennial to seasonal sea ice (Haine and Martin, 2017; see Section 3.2.1.1.2). Other drivers include increased warm, moist air intrusions into the Arctic during both winter (Box 3.1) and spring (Boisvert et al., 2016; Cullather et al., 2016; Kapsch et al., 2016; Mortin et al., 2016; Graham et al., 2017; Hegyi and Taylor, 2018), radiative feedbacks associated with cloudiness and humidity (Kapsch et al., 2013; Pithan and Mauritsen, 2014; Hegyi and Deng, 2016; Morrison et al., 2018), and increased exchanges of sensible and latent heat flux from the ocean to the atmosphere (Serreze et al., 2012; Taylor et al., 2018). A lack of complete process understanding limits a more definitive differentiation between anthropogenic versus internal drivers of summer Arctic sea ice loss (Serreze et al., 2016; Ding et al., 2017; Meehl et al., 2018). The unabated reduction in Arctic summer sea ice since AR5 means contributions to additional global radiative forcing (Flanner et al., 2011) have continued, with estimates of up to an additional 6.4 ± 0.9 W/m^2 of solar energy input to the Arctic Ocean region since 1979 (Pistone et al., 2014).

Although Arctic ice freeze-up is occurring later (Section 3.2.1.1.3), rapid thermodynamic ice growth occurs over thin ice areas after air temperatures drop below freezing in autumn. Later freeze-up also delays snowfall accumulation on sea ice, leading to a thinner and less insulating snowpack (Section 3.2.1.1.6) (Sturm and Massom, 2016). These two negative feedbacks help to mitigate sudden and irreversible loss of Arctic sea ice (Armour et al., 2011).

Total Antarctic sea ice cover exhibits no significant trend over the period of satellite observations (Figure 3.3; 1979 to 2018) (*high confidence*) (Ludescher et al., 2018). A significant positive trend in mean annual ice cover between 1979 and 2015 (Comiso et al., 2017a) has not persisted, due to three consecutive years of below-average ice cover (2016-2018) driven by atmospheric and oceanic forcing (Turner et al., 2017b; Kusahara et al., 2018; Meehl et al., 2019; Wang et al., 2019). The overall Antarctic sea ice extent trend is composed of near-compensating regional changes, with rapid ice loss in the Amundsen and Bellingshausen seas counteracted by rapid ice gain in the Weddell and Ross seas (Holland, 2014) (Figure 3.3). These 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.

Multiple factors contribute to the regionally variable nature of Antarctic sea ice extent trends (Matear et al., 2015; Hobbs et al., 2016b). Sea ice trends are closely related to meridional wind trends (*high confidence*) (Holland and Kwok, 2012; Haumann et al., 2014): poleward wind trends in the Bellingshausen Sea push sea ice closer to the coast (Holland and Kwok, 2012) and advect warm air to the sea ice zone (Kusahara et al., 2017), and the reverse is true over much of the Ross Sea. These meridional wind trends are linked to Pacific variability (Coggins and McDonald, 2015; Meehl et al., 2016; Purich et al., 2016b). 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).

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 (Cross-Chapter Box 7 in Chapter 3), 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*). Because Antarctic sea ice extent has remained below climatological values since 2016, there is still potential for longer-term changes to emerge in the Antarctic (Meehl et al., 2019), similar to the Arctic.

Historical surface observations (Murphy et al., 2014), reconstructions (Abram et al., 2013b), ship records (de la Mare, 2009; Edinburgh and Day, 2016), early satellite images (Gallaher et al., 2014), and model simulations (Gagné et al., 2015) indicate a decrease in overall Antarctic sea ice cover since the early 1960s which is too modest to be separated from natural variability (Hobbs et al., 2016a) (*high confidence*).

3.2.1.1.2 Age and thickness

The proportion of Arctic sea ice at least 5 years old declined from 30% to 2% between 1979 and 2018; over the same period first-year sea ice proportionally increased from approximately 40% to 60–70% (Stroeve and Notz, 2018) (*very likely*) (Sections 3.2.1.1.3 and 3.2.1.1.4). It is *virtually certain* that Arctic sea ice has thinned through volume reductions in satellite altimeter retrievals (Laxon et al., 2013; Kwok, 2018), ocean–sea ice reanalyses (Chevallier et al., 2017) and *in situ* measurements (Renner et al., 2014; Haas et al., 2017). 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). There is emerging evidence that this sea ice volume loss may be unprecedented over the past century (Schweiger et al., 2019). 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 shift to thinner seasonal sea ice contributes to further ice extent reductions through enhanced summer season melt via increased energy absorption (Nicolaus et al., 2012), and it is vulnerable to fragmentation from the passage of intense Arctic cyclones in summer and increased ocean swell conditions (Zhang et al., 2013; Thomson and Rogers, 2014).

Surface 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) (*medium confidence*). Satellite altimeter datasets of Antarctic sea ice thickness are emerging (Paul et al., 2018) but definitive trends are not yet available.

3.2.1.1.3 Seasonality

There is *high confidence* that the Arctic sea ice melt season has extended by 3 days per decade since 1979 due earlier melt onset, and 7 days per decade due to later freeze-up (Stroeve and Notz, 2018). This longer melt season is consistent with the observed loss of sea ice extent and thickness (Sections 3.2.1.1.1; 3.2.1.1.2). While the melt onset trends are smaller, they play a large role in the earlier development of open water (Stroeve et al., 2012b; Serreze et al., 2016) and melt pond development (Perovich and Polashenski, 2012) which enhance the sea ice-albedo feedback (Stroeve et al., 2014b; Liu et al., 2015a). Observed reductions in the duration 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 over 1979-2011 largely followed the spatial pattern of sea ice extent trends with reduced ice cover duration in the Amundsen/Bellingshausen Sea region in summer and autumn owing to earlier retreat and later advance, and increases in the Ross Sea due to later ice retreat and earlier advance (Stammerjohn et al., 2012).

3.2.1.1.4 Motion

Winds associated with the climatological Arctic sea level pressure pattern drive the Beaufort Gyre (Dewey et al., 2018; Meneghello et al., 2018) and the Transpolar Drift Stream (Vihma et al., 2012), which retains sea ice within the central Arctic Basin, and exports sea ice out of the Fram Strait, respectively. There is *high confidence* that sea ice drift speeds have increased since 1979, both within the Arctic Basin and through Fram Strait (Rampal et al., 2009; Krumpen et al., 2019), attributed to thinner ice (Spreen et al., 2011) and changes in wind forcing (Olason and Notz, 2014). Fram Strait sea ice area export estimates range between 600,000 to 1 million km^2 of ice annually, which represents approximately 10% of the ice within the Arctic Basin (*medium confidence*) (Kwok et al., 2013; Krumpen et al., 2016; Smedsrud et al., 2017; Zamani et al., 2019). Sea ice volume flux estimates through Fram Strait are now available from satellite altimeter datasets (Ricker et al., 2018), but they cover too short a time period for robust trend analysis. Observations of extreme Arctic sea ice deformation is attributed to the combination of decreased ice thickness and increased ice motion (Itkin 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 temporally inconsistent satellite records (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'. The few long term surface (auger hole) records of Arctic landfast sea ice thickness all exhibit thinning trends in springtime maximum sea ice thickness since the mid-1960s (*high confidence*): declines of 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 start 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).

Arctic landfast ice is important to northern residents as a platform for travel, hunting, and access to offshore regions (Sections 3.4.3.3, 3.5.2.2). 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 (Box 3.3) (Gearheard et al., 2013). Long term records of Antarctic landfast ice are limited in space and time (Stammerjohn and Maksym, 2016), with a high degree of regional variability in 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 in- and under-ice biota) from 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 across the Arctic (with implications for sea ice ecosystems) as the ice regime shifts to thinner seasonal ice (Olsen et al., 2017; Granskog et al., 2018) (*medium confidence*).

Despite the importance of snow on sea ice (Webster et al., 2018), surface or satellite-derived observations of snowfall over sea ice, and snow depth on sea ice are lacking (Webster et al., 2014). The primary source of snow depth on Arctic sea ice are based on observations collected decades ago (Warren et al., 1999) the utility of which are impacted by the rapid loss of multiyear ice across the central Arctic (Stroeve and Notz, 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) consistent with estimates of higher fractions of liquid precipitation since 2000 (Boisvert et al., 2018) but there is *low confidence* because 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), field measurements (Massom et al., 2001) and ship-based observations (Worby et al., 2008), data are not sufficient in time nor space to assess changes in snow accumulation on Antarctic sea ice.

[START BOX 3.2 HERE]

Box 3.2: Potential for the Polar Cryosphere to Influence Mid-latitude Weather

Since AR5, understanding how observed changes in the Arctic can influence mid-latitude weather has emerged as a societally important topic because hundreds of millions of people can potentially be impacted (Jung et al., 2015). The early to middle part of the Holocene coincided with substantial decreases in net precipitation that may be due to weakening jet stream winds related to Arctic temperatures (Routson et al., 2019). There is only *low* to *medium confidence* in the current nature of Arctic/mid-latitude weather linkages because conclusions of recent analyses are inconsistent (National Research Council, 2014; Barnes and Polvani, 2015; Francis, 2017). The atmosphere interacts with the ocean and cryosphere through radiation, heat, precipitation and wind, but a full understanding of complex inter-connected physical processes is lacking Arctic forcing on the atmosphere from loss of sea ice and terrestrial snow is increasing, but 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, 2018a). Connectivity is reduced by the influence of chaotic internal natural variability and other tropical and oceanic forcing. Part of the scientific disagreement is due to irregular connections in the Arctic to mid-latitude linkage pathways, both within and between years (Overland and Wang, 2018b).

Considerable literature exists on the potential for sea ice loss in the Barents and Kara Seas to drive cold episodes in eastern Asia (Kim et al., 2014; Kretschmer et al., 2016), while sea ice anomalies in the Chukchi Sea and areas west of Greenland are associated with cold events in eastern North America (Kug et al., 2015; Ballinger et al., 2018; Overland and Wang, 2018a). Such connections, however, are only episodic (Cohen et al., 2018). While there is evidence of an increase in the frequency of weak polar vortex events (Screen et al., 2018), studies do not show increases in the number of mid-latitude cold events in observations or model projections (Ayarzaguena and Screen, 2016; Trenary et al., 2016). Potential Arctic/mid-latitude interactions have a more regional tropospheric pathway in November-December (Honda et al., 2009; Chen et al., 2016a; McKenna et al., 2018), whereas January-March has a more hemispheric stratospheric pathway involving migration of the polar vortex off of its usual centred location on the North Pole (Cohen et al., 2012; Nakamura et al., 2016; Zhang et al., 2018b). Overall, changes in the stratospheric polar vortex and Northern Annual Mode are not separable from natural variability, and so cannot be attributed to greenhouse gas forced sea ice loss (Screen et al., 2018).

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 3.2 HERE]

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 (SM3.2.1; Figure SM3.2).

3.2.1.2.1 Temperature

Ocean temperatures and associated heat fluxes have a primary influence on sea ice (e.g., Carmack et al., 2015; Steele and Dickinson, 2016). WGI AR5 (their Section 3.2.2) reported that Canada Basin 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 trends for 1982–2017 reveal summer mixed layer temperatures increasing at about 0.5°C per decade over large sectors of the Arctic basin that are ice-free in summer (Timmermans et al., 2017) (Figure 3.3). This is primarily the result of increased absorption of solar radiation accompanying sea-ice loss (Perovich, 2016). Between 1979 and 2011, the decrease in Arctic Ocean albedo corresponded to more solar energy input to the ocean (*virtually certain*) of approximately 6.4 ± 0.9 Wm^{-2} (Pistone et al., 2014), *likely* reducing the growth of sea ice by up to 25% in both Eurasian and Canadian basins (Timmermans, 2015; Ivanov et al., 2016) (Section 3.2.1.1).

While Atlantic Water Layer temperatures appear to show less variability since 2008, total heat content in this layer continues to increase (Polyakov et al., 2017). Recent changes have been dubbed the 'Atlantification' of the Northern Barents Sea and Eurasian Basin (Arthun et al., 2012; Lind et al., 2018), characterized by weaker stratification and enhanced Atlantic Water Layer heat fluxes further northeast (*medium confidence*). Polyakov et al. (2017) estimate 2-4 times larger heat fluxes in 2014-2015 compared with 2007-2008. In the Canadian Basin, the maximum temperature of the Pacific Water Layer increased by ~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). Over 2001-2014, heat transport associated with Bering Strait inflow increased by

60%, from around 10 TW in 2001 to 16 TW in 2014, due to increases in both volume flux and temperature (Woodgate et al., 2015; Woodgate, 2017) (*low confidence*).

The Southern Ocean is important for the transfer of heat from the atmosphere to the global ocean, including heat from anthropogenic warming (Frölicher et al., 2015; Shi et al., 2018). The Southern Ocean accounted for ~75% of the global ocean uptake of excess heat during 1870-1995 (Figure SM3.2; Frölicher et al., 2015), of which ~43% resided in the Southern Ocean with the remainder redistributed to lower latitudes. Over 1970-2017, observations show that the upper 2000m of the ocean south of 30°S was responsible for 35-43% of the increase in global ocean heat content (Table 3.1). Both models and observations show that, relative to its size (Table SM3.1), the Southern Ocean is disproportionately important in the increase in global upper ocean heat content (*high confidence*). Multi-decadal warming of the Southern Ocean has been attributed to anthropogenic factors, especially the role of greenhouse gases but also ozone depletion (Armour et al., 2016; Shi et al., 2018; Swart et al., 2018; Irving et al., 2019) (*medium confidence*).

Table 3.1.: Ocean heat content trend (0-2000m depth) during 2005-2017 and 1970-2017 for the global ocean and Southern Ocean. Ordinary Least Square (OLS) method is used; units are 10^{21} J yr^{-1}. Uncertainties denote the 90% confidence interval accounting for the reduction in the degrees of freedom implied by temporal correlations of residuals, as per Section 5.2. Values in curved brackets are percentages of heat gain by the Southern Ocean relative to the global ocean. Data sources are as per Table SM3.1. The mean proportion and its 5%-95% confidence interval (1.65 times standard deviation of individual estimates) are in the last column.

OHC Trend (10^{21} J yr^{-1})	**Ishii V7.2**	**IAP**	**EN4-GR10**	**IPRC**	**Scripps**	**JAMSTEC**	**Mean [5%, 95%]**
Global 2005-17	10.06±1.28	8.45±1.04	10.57±1.17	9.96±1.57	8.38±1.31	9.06±0.67	
South of 30°S 2005-17	5.20±1.03 (52%)	4.55±1.00 (54%)	5.38±1.30 (51%)	6.24±1.80 (63%)	4.22±0.70 (50%)	4.44±0.63 (49%)	**53% [45%, 62%]**
Global 1970-2017	6.73±0.55	7.02±1.96	5.28±1.01				
South of 30°S 1970-2017	2.42±0.26 (36%)	2.78±0.29 (40%)	2.18±0.36 (41%)				**39% [35%, 43%]**

Surface warming during 1982-2016 was strongest along the northern flank of the Antarctic Circumpolar Current (ACC), contrasting with cooling further south (Figure 3.3). Interior warming was strongest in the upper 2000 m, peaking around 40-50°S (Armour et al., 2016) (SM3.2.1; Figures SM3.2 and SM3.3). There is *high confidence* that this pattern of change is driven by upper-ocean overturning circulation and mixing (Cross-Chapter Box 7 in Chapter 3), 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 strongly influenced by wind-forced changes in the thickness and depth of the mode water layer (Desbruyeres et al., 2017; Gao et al., 2018) (*medium confidence)*. Below the surface south of the ACC, warming extends close to Antarctica, intruding onto the continental shelf in the Amundsen-Bellingshausen Sea where temperature increases of 0.1-0.3°C per decade have been observed over 1983-2012 (Schmidtko et al., 2014) (Section 3.3.1.5). This latter warming may be driven by changes in wind forcing (Spence et al., 2014), and exhibits significant decadal variability (Jenkins et al., 2018).

After around 2005, improved upper ocean heat content estimates became available via Argo profiling floats (Section 1.8.1; Section 5.2). For 2005-2017, multiple datasets show that the heat gained by the Southern Ocean south of 30°S was 45-62% of the global ocean heat gain (Table 3.1) (equivalent figures for other indicative Southern Ocean extents are in Table SM3.2). This accords with Roemmich et al. (2015), who found that during 2006-2013 the ocean south of 20°S accounted for 67-98% of total heat gain in the upper 2000 m of the global ocean. (The smaller proportion for 2005-2017 c.f. 2006-2013 is due to comparatively greater warming in the earlier part of the common period). The recent Southern Ocean heat gain is thus larger than its long-term trend over either the preceding several decades (1970-2004, 30-51%, Table SM3.3) or the full period 1970-2017 (35-43%; Table 3.1 and above). There is *high confidence* that the Southern Ocean has increased its role in global ocean heat content in recent years compared with the past several decades. Attribution of this increased role is currently lacking.

The ocean below 2000 m globally stores ~19% of the excess anthropogenic heat in the Earth system, with a large fraction (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 WGI 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). Antarctic Bottom Water volume is decreasing (Purkey and Johnson, 2012), resulting in a deepening of density surfaces and driving much of the warming on depth surfaces below 2000 m (Desbruyeres et al., 2017). This reduction in bottom water volume is suggestive of a decrease in its production (Purkey and Johnson, 2013). In the Indian and Pacific basins close to Antarctica, bottom water is freshening (Purkey and Johnson, 2013; Menezes et al., 2017) consistent with the uptake of enhanced Antarctic ice-shelf and glacial melt (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. Salinity changes are induced by freshwater runoff to the ocean (rivers and land ice), net precipitation, sea ice, and advection of mid-latitude waters, with the potential to impact water mass formation and circulation (e.g., Thornalley et al., 2018; see also Section 6.7.1).

Updating WGI AR5 (their Section 3.3.3.3), recent Arctic-wide estimates yield a freshwater increase (relative to salinity of 34.8 on the Practical Salinity Scale, used throughout this chapter) of 600 ± 300 km^3 yr^{-1} over 1992-2012, with about two-thirds concomitant with decreasing salinity, and the remainder with a thickening of the freshwater layer (*medium confidence*) (Rabe et al., 2014; Haine et al., 2015; Carmack et al., 2016). The Beaufort Gyre region has increased its freshwater by ~40% (6,600 km^3) over 2003-2017; this, and the Gyre's strengthening, have been attributed to dominance of clockwise wind patterns over the Canadian Basin over 1997-2016 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 ~180 km^3 over 2003-2014 (Armitage et al., 2016). During the 2000s, freshwater content in the upper 100 m of the northern Barents Sea declined by about 32%, from a mean of ~2.5 m (relative to a salinity of 35) in 1970-1999, to 1.7 m in 2010-2016 (Lind et al., 2018). An increasing trend of 30 ± 20 km^3yr^{-1} in freshwater flux through Bering Strait, primarily due to increased volume flux, was measured from 1991-2015, with record maximum freshwater influx in 2014 of around 3,500 km^3 in that year (Woodgate, 2017). Freshwater fluxes from rivers are also increasing (Section 3.4.1.2.2), and there have been observed increases in discharge of glacial ice from Greenland (Section 3.3.1.3).

Observed Southern Ocean freshening trends are consistent with WGI AR5; subsequent studies have increased confidence in their magnitude and sign, and also attributed them to anthropogenic influences (Swart et al., 2018). Changes over 1950–2010 show persistent surface water freshening over the whole Southern Ocean, with subducted mode/intermediate waters carrying trends of 0.0002–0.0008 yr^{-1} to below 1500 m (Skliris et al., 2014), whilst de Lavergne et al. (2014) observe a circumpolar freshening south of the ACC of 0.0011 ± 0.0004 yr^{-1} in the upper 100 m since the 1960s (*medium confidence*). This intensifies over the Antarctic continental shelves (except along the Western Antarctic Peninsula), where freshening of up to 0.01 yr^{-1} is observed (Schmidtko et al., 2014). Freshening may be driven by increases in precipitation, but while models (Pauling et al., 2016) and observations suggest an increase may have occurred over the last 60 years, uncertainty is presently too high to quantify its net impact (Skliris 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% from 1982–2008, and that this may have driven freshening of 0.002 ± 0.001 yr^{-1} in the surface and intermediate waters. 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 (Cross-Chapter Box 7 in Chapter 3). Freshwater input to the ocean from the Antarctic Ice Sheet also has the potential to affect the properties and circulation of Southern Ocean water masses; see Section 3.3.3.

3.2.1.2.3 Stratification

See Supplementary Material (SM3.2.2).

3.2.1.2.4 Carbon and ocean acidification

Various elements of marine biogeochemistry and geochemistry in the polar regions are of global importance. Here we focus on aspects relevant to carbon and ocean acidification; others (e.g. changes in dissolved oxygen) are assessed in Section 5.2.2.

About a quarter of carbon dioxide (CO_2) released by human activities is taken up by the ocean (WGI AR5). This dissolves in surface water to form carbonic acid, which, upon dissociation, causes a decrease in pH (acidification) and carbonate ion (CO_3^{2-}) concentration. This can affect organisms that form shells and skeletons using calcium carbonate ($CaCO_3$, aragonite and calcite as dominant mineral forms). 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).

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 ~100 m between the 1990s and 2010 primarily due to increased Pacific Winter Water transport. In the East Siberian Arctic Shelf, extreme aragonite undersaturation was driven by the degradation of terrestrial organic matter and runoff of Arctic river water with elevated CO_2 concentrations, reflecting pH changes in excess of those projected in this region for 2100 (Semiletov et al., 2016) (*high confidence*); this was also observed along the continental margin and traced in the deep Makarov and Canada Basins (Anderson et al., 2017a). The variable buffering capacities of rivers flowing through watersheds with different bedrock geology also influenced the state of ocean acidification in coastal regions (Tank et al., 2012; Azetsu-Scott et al., 2014).

The dissolved inorganic carbon (DIC) concentration increased in subsurface waters (150-1400m) in the central Arctic between 1991 and 2011 (Ericson et al., 2014). 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 waters below 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 TgC yr^{-1} of DIC across the Arctic Ocean gateways to the North Atlantic; at least 166 ± 60 TgC yr^{-1} of this was sequestered from the atmosphere (*medium confidence)*.

Studies covering seasonal-to-decadal variability in the Arctic are limited, with most conducted in ice-free or low-ice periods during summer-autumn. However, it has been demonstrated that biological processes, respiration and photosynthesis, control the $CaCO_3$ saturation states in Chukchi Sea bottom water (Yamamoto-Kawai et al., 2016). Sea ice formation and melt influence the dynamics of ikaite ($CaCO_3$ precipitation trapped in sea-ice during brine rejection), and therefore local carbonate chemistry (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 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 sustain high pH in kelp forests, slowing ocean acidification (Krause-Jensen et al., 2016).

Since AR5, there are new constraints on the seasonal-to-decadal variability in the Southern Ocean CO_2 flux (McNeil and Matear, 2013; Landschützer et al., 2014; Landschützer et al., 2015; Gregor et al., 2017; Ritter et al., 2017; Keppler and Landschutzer, 2019) (Figure SM3.4), with mean annual flux anomalies varying from 0.3± 0.1 Pg C yr^{-1} in 2001-2002 to -0.4 Pg C yr^{-1} in 2012 (Landschützer et al., 2015); this can affect the magnitude of the global CO_2 sink (Section 5.2.2). A weakening CO_2 sink during the 1990s (Le Quéré et al., 2007) reversed in the 2000s as part of a decadal cycle (Landschützer et al., 2015; Munro et al., 2015; Williams et al., 2017) (SM3.2.3; Figure SM3.4), with a weakening again since 2011 (Keppler and Landschutzer, 2019). While the weakening sink during the 1990s was explained as a response to changes in the circumpolar winds over the Southern Ocean enhancing the outgassing of natural CO_2, the subsequent changes appear due to a combination of changes in regional winds, temperature, and circulation (Landschützer et al., 2015; Gregor et al., 2017; Keppler and Landschutzer, 2019). Data scarcity, especially in winter, remains a challenge (Gruber et al., 2017; Ritter et al., 2017; Fay et al., 2018); recent data from pH-enabled floats highlighted the potential role for winter outgassing south of the Polar Front (Williams et al., 2017; Gray et al., 2018). Overall, there is *medium confidence* that the Southern Ocean CO_2 sink has experienced significant decadal variations since the 1980s.

Southern Ocean carbon storage is affected by changes in overturning circulation (Cross-Chapter Box 7 in Chapter 3), with the storage of anthropogenic and natural carbon being both variable and out of phase on decadal timescales (DeVries et al., 2017; Tanhua et al., 2017) (Table SM3.4). Mode and intermediate waters are strongly involved in changing storage, also showing high sensitivity to shifts in winds (Swart et al., 2014; Swart et al., 2015a; Tanhua et al., 2017; Gruber et al., 2019). Zonal basin differences in the uptake and storage of anthropogenic carbon are not well resolved and there is weak agreement between reanalysis products and CMIP5 models (Swart et al., 2014). The presence of subduction hotspots suggest that basin-wide studies may be underestimating the importance of mode water subduction as a principal storage mechanism (Langlais et al., 2017).

Strengthening impacts of Southern Ocean acidification are illustrated by the 3.9 ± 1.3% decrease in derived calcification rates (1998–2014) (Freeman and Lovenduski, 2015). These have strong regional character, with decreases in the Indian and Pacific sectors (7.5–11.6%) and increases in the Atlantic (14.3 ± 5.1%). There have also been changes in the seasonality of pCO_2 linked to decreasing buffer capacity (McNeil and Sasse, 2016) (SM3.2.4) or adjustments to primary production (Conrad and Lovenduski, 2015); seasonal changes are discussed further in Section 5.2.2.

3.2.1.3 Ocean Circulation

The major elements of Southern Ocean circulation are assessed in Cross-Chapter Box 7 in Chapter 3; Arctic Ocean circulation is considered here. Arctic processes, such as discharge of freshwater from the Greenland Ice Sheet, have the potential to impact on the formation of the headwaters of the Atlantic Meridional Overturning Circulation (Section 6.7.1), and can impact on the structure and function 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 some currents in the Beaufort Gyre approximately doubled (Armitage et al., 2017). Over 2001-2014, annual Bering Strait volume transport from the Pacific to the Arctic Ocean increased from $0.7x10^6$ m^3s^{-1} to $1.2x10^6$ m^3s^{-1} (Woodgate et al., 2015). Mesoscale eddies are characterized by horizontal scales of ~10 km in the Arctic, and are important components of the ocean system. 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 this is *about as likely as not* (Meneghello et al., 2017). Data of sufficiently high resolution 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 (Cross-Chapter Box 7 in Chapter 3), there is comparatively little knowledge on changing Arctic frontal positions and current cores since AR5. An exception is that the Beaufort Gyre expanded to the northwest between 2003 and 2014, contemporaneous with changes in its freshwater accumulation and alterations in wind forcing, resulting in increased proximity to the Chukchi Plateau and Mendeleev Ridge (Armitage et al., 2017; Regan et al., 2019) (Section 3.2.1.2.2).

[START CROSS-CHAPTER BOX 7 HERE]

Cross-Chapter Box 7: Southern Ocean Circulation: Drivers, Changes and Implications

Authors: Michael P. Meredith (UK), Robert Hallberg (US), Alessandro Tagliabue (UK), Andrew Meijers (UK/Australia), Jamie Oliver (UK), Andrew Hogg (Australia)

Horizontal circulation and movement of fronts

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 horizontal circulation in the circumpolar Southern Ocean is comprised of an eastward-flowing mean current concentrated in a series of sinuous, braided jets exhibiting strong meandering variability and shedding small-scale transient eddies (Figure CB7.1). The mean flow circumnavigates Antarctica as the world's largest ocean current, the Antarctic Circumpolar Current (ACC), transporting approximately $173.3 \pm 10.7 \times 10^6$ m^3 s^{-1} (Donohue et al., 2016) of water eastward in a geostrophic balance set up by the contrasting properties of waters around Antarctica and those inside the subtropical gyres to the north of ACC. This contrast is maintained by a combination of strong westerly winds and ocean heat loss south of the ACC.

Trends in the atmospheric forcing of the Southern Ocean are dominated by a strengthening of westerly winds in recent decades (Swart et al., 2015a), but there is no evidence that this enhanced wind stress has significantly altered the ACC transport. While the annual mean value of transport is stable in the instrumental period (Chidichimo et al., 2014; Koenig et al., 2014; Donohue et al., 2016) it is difficult to resolve changes in barotropic transport; overall there is *medium confidence* that ACC transport is only weakly sensitive to changes in winds. This is consistent with longer-term analyses that find only minimal changes in ACC transport since the last glaciation (McCave et al., 2013). Theoretical predictions and high-resolution ocean modelling suggest that the weak sensitivity of the ACC to changes in wind stress is a consequence of eddy saturation (Munday et al., 2013), whereby the time-mean state of the ocean remains close to a marginal condition for eddy instability and hence 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, 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).

WGI 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; Meijers et al., 2019) (Section 3.2.3.2). Since AR5, however, substantial contrary evidence has emerged. While winds have strengthened over the Southern Ocean, reanalysis products show no significant shift in the annual mean latitude of zonal wind jets between 1979-2009 (Swart et al., 2015a). Similarly, 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 WGI 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 increases in sea surface height consistent with mean global trends in sea level rise (Gille, 2014). The increase in sea surface height is ascribed largely to warming-driven steric expansion in the upper ocean, but the mechanism driving such warming is still uncertain (Gille, 2014). These recent findings do not preclude more local changes in frontal position, but it is now assessed as *unlikely* that there has been a statistically significant net southward movement of the mean ACC position over the past 20 years.

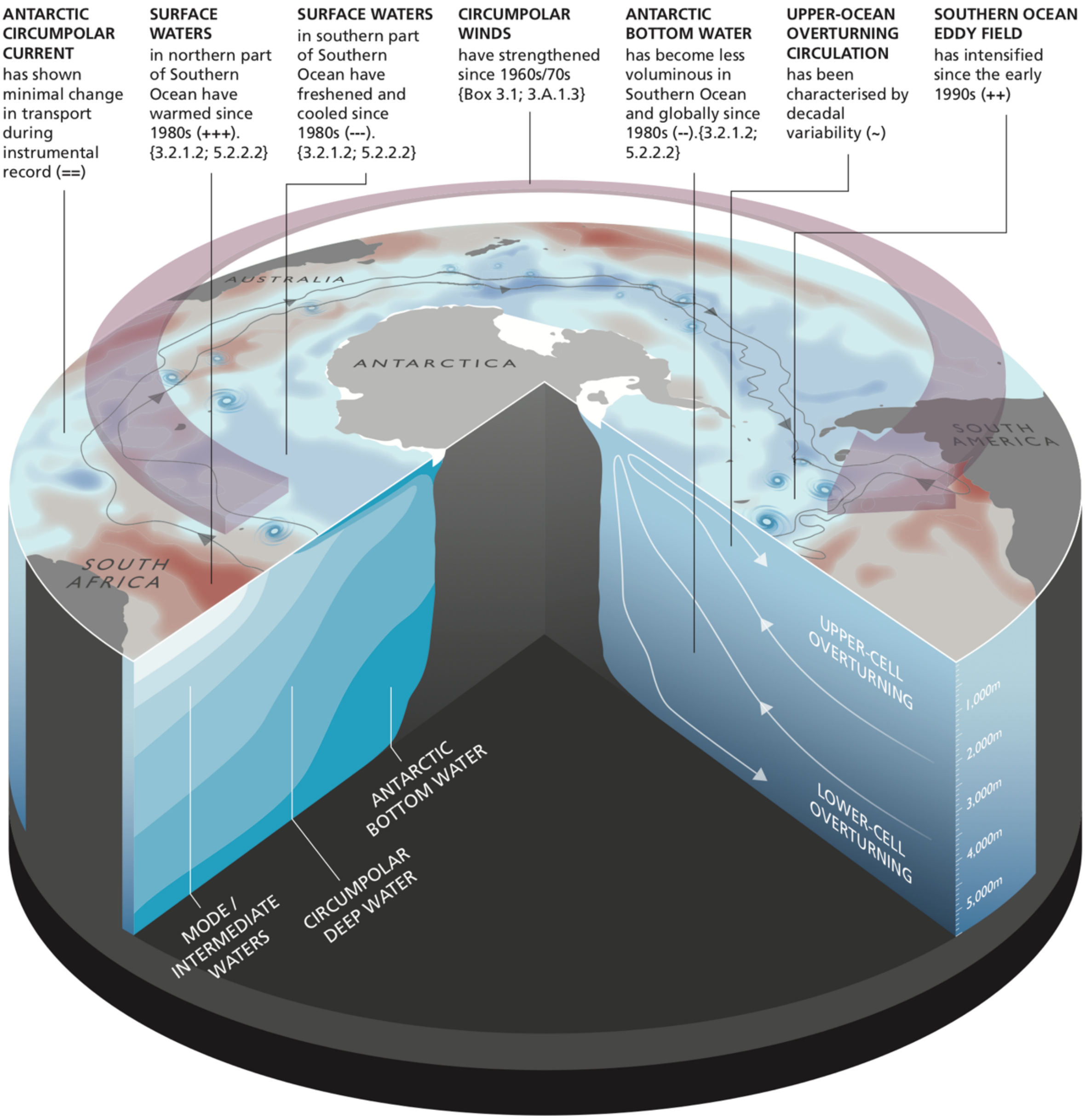


Figure CB7.1: Schematic of some of the major Southern Ocean changes assessed in this Box and in Chapters 3 and 5. Assessed changes are marked as positive (+), neutral (=), negative (-), or dominated by variability (~). The number of symbols used indicates confidence, from *low* (1) through *medium* (2) to *high* (3). Section numbers indicate the links to further information outside this box.

Overturning circulation and water mass formation

The Southern Ocean is 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. Such export of both extremely cold and dense Antarctic Bottom Water and the lighter mode and intermediate waters (Figure CB7.1) represents important pathways for surface properties to be sequestered from the atmosphere for decades to millennia. This upwelling and sinking constitutes a two-limbed overturning circulation, by which much of the global deep ocean is renewed.

The 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 sequesters nutrients that ultimately end up supporting 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 (Section 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).

It is challenging to measure the Southern Ocean overturning directly, and misinterpretation of Waugh et al. (2013) led to AR5 erroneously reporting the upper cell to have slowed (AR5 WGI, Section 3.6.4). However, additional 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-2000s water mass ages changed in a manner consistent with an increase in upwelling and overturning. However, inverse analyses suggest that such overturning experiences significant inter-decadal variability in response to wind forcing, with reductions in 2000-2010 relative to 1990-2000 (DeVries et al., 2017). This variability, combined with the indirect nature of observational estimates, means that there is *low confidence* in assessments of long-term changes in upper cell 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); recently-updated analyses confirm this trend to present day (Figure 5.4). This suggests that the production and export of this water mass has probably slowed, though direct observational evidence is difficult to obtain. 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).

Projections

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 and changes in sea ice export (Downes and Hogg, 2013). Dynamical considerations and numerical simulations indicate that, if further increases in the westerly winds are sustained, then it is *very likely* that the eddy field will continue to grow in intensity (Morrison and McC. Hogg, 2013; Munday et al., 2013), with potential consequences for the upper-ocean overturning circulation and transport of tracers (Abernathey and Ferreira, 2015) (including heat, carbon, oxygen and nutrients), and *likely* that the mean position and strength of the ACC will remain only weakly sensitive to winds.

The considerable CMIP5 inter-model variations in Southern Ocean time-mean circulation projections reported in WGI AR5 (Meijers et al., 2012; Downes and Hogg, 2013) remain largely unchanged. Some of the differences in projected changes have been found to be correlated with biases in the various models' ability to simulate the historical state of the Southern Ocean, such as mixed layer depth (Sallée et al., 2013a) and westerly wind jet latitude (Bracegirdle et al., 2013). This suggests that bias reduction against observed historical metrics (Russell et al., 2018) in future generations of coupled models (e.g., CMIP6) should lead to improved confidence in aspects of projected Southern Ocean changes. CMIP5 models suggest that the subduction of mode and intermediate water will increase (Sallée et al., 2013b), which will affect oxygen and nutrient transports, and the overall transport of the Southern Ocean upper overturning cell will increase by up to 20% (Downes and Hogg, 2013), but model performance is limited by the inability to explicitly resolve 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 global models. Further uncertainty derives from increased meltwater from the Antarctic Ice Sheet not being considered in the CMIP5 climate models, despite its potential for significant impact on Southern Ocean dynamics and the global climate, and its potential for positive feedbacks (Bronselaer et al., 2018). Due to these uncertainties, *low confidence* is therefore ascribed to the CMIP5-based model projections of future Southern Ocean circulation and water masses.

[END CROSS-CHAPTER BOX 7 HERE]

3.2.2 Projected Changes in Sea Ice and Ocean

3.2.2.1 Sea Ice

The multi-model ensemble of historical simulations from CMIP5 models identify declines in total Arctic sea ice extent and thickness (Sections 3.2.1.1.1; 3.2.1.1.2; Figure 3.3) which agree with observations (Massonnet et al., 2012; Stroeve et al., 2012a; Stroeve et al., 2014a; Stroeve and Notz, 2015). There is a range in the ability of individual models to simulate observed sea ice thickness spatial patterns and sea ice drift rates (Jahn et al., 2012; Stroeve et al., 2014a; Tandon et al., 2018). Reductions in Arctic sea ice extent scale linearly with both global temperatures and cumulative CO_2 emissions in simulations and observations (Notz and Stroeve, 2016), although aerosols influenced historical sea ice trends (Gagné et al., 2017). The uncertainty in sea ice sensitivity (ice extent loss per unit of warming) is quite large (Niederdrenk and Notz, 2018) and the model sensitivity is too low in most CMIP5 models (Rosenblum and Eisenman, 2017). Emerging evidence suggests, however, that internal variability, including links between the Arctic and lower latitude, strongly influences the ability of models to simulate observed reductions in Arctic sea ice extent (Swart et al., 2015b; Ding et al., 2018).

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 (Massonnet et al., 2012; Stroeve et al., 2012a; Overland and Wang, 2013) as a result of natural climate variability (Notz, 2015; Swart et al., 2015b; Screen and Deser, 2019), scenario uncertainty (Stroeve et al., 2012a; 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 results in an uncertainty of approximately 20 years in the timing of seasonally ice-free conditions (Notz, 2015; Jahn, 2018), but the clear link between summer sea ice extent and cumulative CO_2 emissions provide a basis for when consistent ice-free conditions may be expected. For stabilized global warming of 1.5°C, sea ice in September is *likely* to be present at end of century with an approximately 1% chance of individual ice-free years (Notz and Stroeve, 2016; Sanderson et al., 2017; Jahn, 2018; Sigmond et al., 2018); after 10 years of stabilized warming at a 2°C increase, more frequent occurrence of an ice-free summer Arctic is expected (around 10-35%) (Mahlstein and Knutti, 2012; Jahn et al., 2016; Notz and Stroeve, 2016). Model simulations show that a temporary temperature overshoot of a 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., 2013; Shu et al., 2015). The ensemble mean across multiple models show a decrease in total Antarctic sea ice extent during the satellite era, in contrast to the lack of any observed trend (Figure 3.3; Section 3.2.1.1.1). Interannual sea ice variability in the models is 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. During the historical period, 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). 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).

There is *low confidence* in projections of Antarctic sea ice because there are multiple anthropogenic forcings (ozone and greenhouse gases) and complicated processes involving the ocean, atmosphere, and adjacent ice sheet (Section 3.2.1.1.). Model deficiencies are related to stratification (Sallée et al., 2013a), freshening by ice shelf melt water (Bintanja et al., 2015), atmospheric processes including clouds (Schneider and Reusch, 2015; Hyder et al., 2018), and wind and ocean-driven processes (Purich et al., 2016a; Purich et al., 2016b; Schroeter et al., 2017; Purich et al., 2018; Zhang et al., 2018a). Uncertainty in sea ice projections reduces confidence in projections of Antarctic Ice Sheet surface mass balance because sea ice affects Antarctic temperature and precipitation trends (Bracegirdle et al., 2015), and impacts projected changes in the

Southern Hemisphere westerly jet (Bracegirdle et al., 2018; England et al., 2018) with implications for the Southern Ocean overturning circulation (Cross-Chapter Box 7 in Chapter 3).

[START BOX 3.3 HERE]

Box 3.3: Polynyas

Arctic Coastal Polynyas

Arctic polynyas (areas of open water surrounded by sea ice) are important because they ventilate the Arctic Ocean. The polynyas induce bottom reaching convection on shallow shelves (Damm et al., 2018) because the warm and exposed ocean surface creates very high heat fluxes and new sea ice formation during winter, releasing brine and creating dense water (Barber et al., 2012). On the shallow Siberian shelves, the ocean surface waters are dominated by river runoff which is rich with sediments (Damm et al., 2018), which end up both in the dense bottom water and in new sea ice (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 ice-formation processes. Further reductions in sea ice are projected for Arctic shelf seas which have already lost ice in recent decades (Barnhart et al., 2015; Onarheim et al., 2018) so polynyas will cease to exist where seasonal sea ice disappears or evolve to become part of the marginal sea ice zone due to changes in ice dynamics (i.e., the North Water polynya and the Circumpolar Flaw Lead); 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 (*medium confidence*). The reduced survival rate of sea-ice in the Transpolar Drift interrupts the transport of sediment-laden ice produced from Siberian shelf polynyas (Krumpen et al., 2019), with consequences for the associated biogeochemical matter and gas fluxes (Damm et al., 2018) (*medium confidence*).

Projected changes to polynyas are important because the spring phytoplankton bloom starts early as the ocean is often well-ventilated and nutrient rich, so the entire biological range from phytoplankton to seabirds to marine mammals thrive in polynya waters (*high confidence*) (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, Arctic peoples have hunted regularly in these areas for 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 other 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 (Cross-Chapter Box 3 in Chapter 1).

Antarctic Coastal Polynyas

The Antarctic continent is surrounded by coastal polynyas, which form from the combined effects of winds and landfast ice in the lee of coastal features that protrude into the westward coastal current (Nihashi and Ohshima, 2015; Tamura et al., 2016). 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 Ross and Weddell sea polynyas and around East Antarctica (Drucker et al., 2011; Nihashi and Ohshima, 2015; Tamura et al., 2016) (*high confidence*).

Antarctic coastal polynyas are biological hot-spots that support high rates of primary production (Ainley et al., 2015; Arrigo et al., 2015) due to a combination of both high light (Park et al., 2017) and high nutrient levels, especially iron (Gerringa et al., 2015). Basal ice shelf melt is the primary supplier of iron to coastal polynyas (Arrigo and van Dijken, 2015) although sea ice melt and intrusions of Circumpolar Deep Water are significant in the Ross Sea (McGillicuddy et al., 2015; Hatta et al., 2017). 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; Malpress et al., 2017) (Section 3.2.3.2.4). 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 re-mineralized 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, there is evidence that future changes in ice shelf melt rates could increase water column productivity (Gerringa et al., 2015; Rickard and Behrens, 2016; Kaufman et al., 2017), 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 close to the Maud Rise seamount (at approximately 65ºS, 3ºE), and has importance on a global scale for deep water ventilation. The polynya opens intermittently, and remained open from 1974 to 1976, with an area of 0.2–0.3 million km^2 (Carsey, 1980). A similar polynya appeared in spring 2017, with a smaller area in 2016, but did not occur in 2018 (Campbell et al., 2019; Jena et al., 2019). Based on these recent events, there is *medium confidence* in the drivers of Weddell Polynya formation - it forms over deep water and appears connected to sea ice divergence created by ocean eddies (Holland, 2001) or strong winds (Campbell et al., 2019; Francis et al., 2019; Wilson et al., 2019). Around Maud Rise, the ocean is weakly stratified, and winter sea ice formation causes brine release and the related deepening mixed layer brings warmer deep waters towards the surface. This causes heat loss to the atmosphere above 200 W m^{-2} (Campbell et al., 2019). These polynya formation processes 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., 2017). 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 resulting in *low confidence* in future Weddell Polynya projections (Reintges et al., 2017).

[END BOX 3.3 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 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 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 were 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 simulations for RCP8.5, Nummelin et al. (2017) found a 2–6°C range in Arctic amplification of surface air temperature north of 70°N, consistent with increased ocean heat transport. Comparing 26 different CMIP5 simulations 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 appears as a probable mechanism with continued warming (Koenigk and Brodeau, 2014; Årthun et al., 2019). Based on 4 CMIP5 models, the Barents Sea is projected to become 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). The ocean heat transport increases in all Arctic gateways, but is dominated by the Barents Sea, and when winter sea ice disappears here the heat loss cannot increase further and the excess ocean heat continues into the Arctic Basin (Koenigk and Brodeau, 2014).

The surface mixed layer of the Arctic Ocean is expected to freshen in 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 configurations comparable to CMIP5, suggesting limited predictive skill for the Arctic freshwater cycle.

CMIP5 projections (Figure 3.3) indicate that observed Southern Ocean warming trends will continue under RCP4.5 and RCP8.5 scenarios, leading to 1–3°C warming by 2100 mostly in the upper ocean (Sallée et al., 2013a). 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 also becomes considerably fresher (salinity decrease of approximately 0.1) (Sallée et al., 2013b) with an overall increase in stratification and a shallowing of mixed layers (Sallée et al., 2013a). 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., 2013a) 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 7 in Chapter 3.

3.2.2.3 Carbon and Ocean Acidification

The Arctic and Southern Ocean have a systemic vulnerability to aragonite undersaturation (Orr et al., 2005). For the RCP8.5 scenario, the entire Arctic and Southern Ocean surface waters will *very likely* be typified by year-around conditions corrosive for aragonite minerals for 2090-2100 (Figure 3.4) (Hauri et al., 2015; Sasse et al., 2015), whilst under RCP2.6 the extent of undersaturated waters are reduced markedly. At a basin/circumpolar scale, there is *high confidence* in these projections due to our robust understanding of the driving mechanisms. However, there is *medium confidence* for the response of specific locations, due to the need for improved resolution of the local circulation, interactions with sea ice, and other processes that modulate the rate of acidification.

Under RCP8.5, melting ice causes the greatest declining rate of pH and $CaCO_3$ saturation state in the Central Arctic, Canadian Arctic Archipelago and Baffin Bay (Popova et al., 2014). In the Canada Basin, projections using RCP8.5 show reductions in mean surface pH from approximately 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, a large increase in area extent of undersaturated surface waters, and a pH change in the surface water of −0.19 are projected using the SRES A1B scenario (broadly comparable to RCP6.0) in the Nordic Sea from 2000 to 2065 (Skogen et al., 2014). Under the same scenario, aragonite undersaturation is projected to occur in the bottom waters over the entire Kara Sea shelf by 2040 and over most of the Barents and East Greenland shelves by 2070 due to the accumulation of anthropogenic CO_2 (Wallhead et al., 2017).

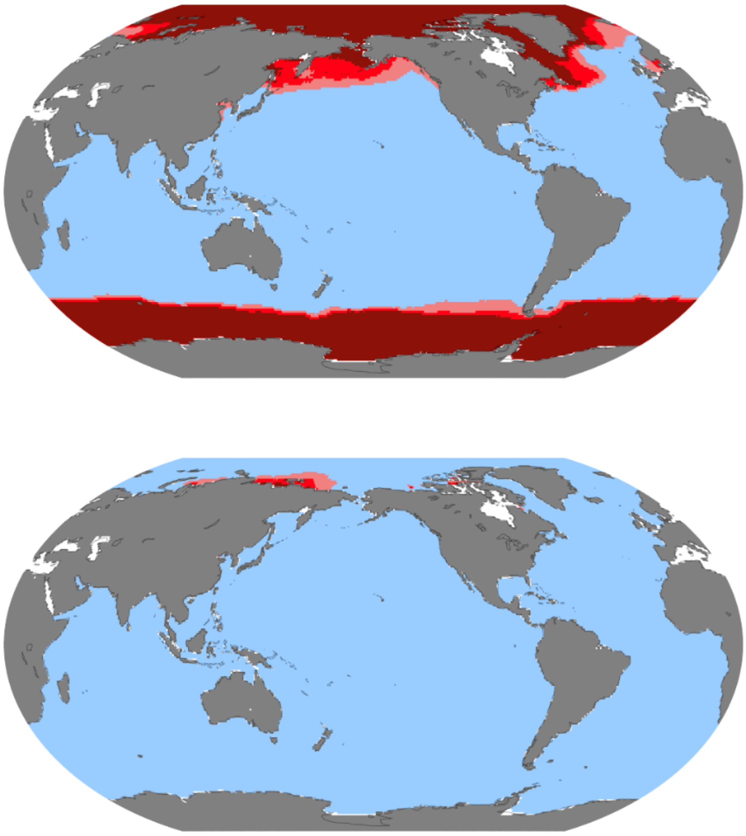


Figure 3.4: The upper ocean (0-10m) at end of this century (2081-2100), characterised by undersaturated conditions for aragonite across 90% confidence intervals (dark red to light red) for the RCP8.5 (top) and RCP2.6 (bottom) scenarios in CMIP5. Saturation states are averaged and confidence intervals calculated at each geographic location across the CNRM-CM5, HadGEM2-ES, GFDL-ESM2G, GFDL-ESM2G, IPSL-CM5-LR, IPSL-CM5-MR, MPI-LR, MPI-MR and NCAR-CESM1 models

Under RCP8.5, the rate of CO_2 uptake by the Southern Ocean is projected to 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 rate will slow and likely stop around 2070±10 corresponding to cumulative CO_2 emissions of 1600 Gt C (Kessler and Tjiputra, 2016; Wang et al., 2016b). This halt in the increase in the uptake rate of CO_2 is linked to the combined feedbacks from well-understood reductions in buffering capacity and warming, as well as the increased upwelling rate of carbon-rich Circumpolar Deep Water (Hauck and Volker, 2015) (Cross-Chapter Box 7 in Chapter 3). Although there is *high agreement* amongst models, contemporary biases in the fluxes of CO_2 in CMIP5 models in the Southern Ocean (Mongwe et al., 2018) suggest *medium confidence* levels for these projections.

Alongside the mean state changes, Southern Ocean aragonite saturation is also affected by the seasonal cycle of carbonate as well as by the impact of reduced buffering capacity (SM3.2.4) on the seasonal cycle of CO_2 (Sasse et al., 2015; McNeil and Sasse, 2016). This leads to an amplification of the seasonal variability of pCO_2 and the hydrogen ion concentration (Hauck and Volker, 2015; McNeil and Sasse, 2016; Landschützer et al., 2018) that accelerates the onset of hypercapnia (i.e., high pCO_2 levels; pCO2 > 1000 µatm) to nearly 2 decades (~2085) ahead of anthropogenic CO_2 forcing (McNeil and Sasse, 2016). The seasonal cycles of pH and aragonite saturation will be attenuated (Kwiatkowski and Orr, 2018) (Section 5.2.2.3), however when the mean state changes are combined with the changes in seasonality, the onset of undersaturation is brought forward by 10-20 years (Table SM3.5). Model projections remain uncertain and affected by the resolution of local ocean physics, which leads to overall *medium confidence* in the timing of undersaturation and hypercapnia.

3.2.3 Impacts on Marine Ecosystems

3.2.3.1 Arctic

Climate change has, and is projected to continue to have, significant implications for Arctic marine ecosystems, with consequences at different trophic levels both in the pelagic, benthic, and sympagic (sea ice related) realms (Figure 3.5). Specifically, climate change is projected to alter the distribution and properties of Arctic marine habitats with associated implications for species composition, production and ecosystem structure and function (Frainer et al., 2017; Kaartvedt and Titelman, 2018; Moore et al., 2018). The rate and severity of ecosystem impacts will be spatially heterogeneous and dependent on future emission scenarios.

In the few Arctic regions where data is sufficient to assess trends in biodiversity, the ecosystem level responses appear to be products of multiple interacting physical, chemical and biological processes (Frederiksen, 2017) (*medium confidence*). Climate change impacts on vertical fluxes and stratification (Sections 3.2.1.2.3, 3.2.2.2) will contribute to changes in bentho-pelagic-sympagic coupling. For instance, projected climate driven changes in ocean properties and hydrography (Section 3.2.2.2) and the abundance of pelagic grazers (Box 3.4) could alter the export of organic matter to the sea floor with associated impacts on the benthos in some Arctic shelf ecosystems (Moore and Stabeno, 2015; Stasko et al., 2018) (*low confidence*). Projected future reductions in summer sea ice (Section 3.2.1.1), increased stratification in summer, shifting currents and fronts and increased ocean temperatures (Section 3.2.2.2) and ocean acidification (Section 3.2.2.3) are all expected to impact the future production and distribution of several marine fish and invertebrates (*high confidence*).

Ocean acidification (Section 3.2.2.3) will affect several key Arctic species (*medium confidence*). The effects of current and projected levels of acidification have been examined for a broad suite of species groups (bivalves, cephalopods, echinoderms, crustaceans, corals and fishes) and these studies reveal species-specific differences in sensitivity, as well as differences in the scope for, and energetic cost of, adaptation (Luckman et al., 2014; Howes et al., 2015; Falkenberg et al., 2018).

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 primary producers (Moore et al., 2018) (*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*). Blooms of this size (1000s of km^2) and intensity (peaks of approximately 30 mg Chla-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 (Arrigo et al., 2012). Evidence shows that these blooms can thrive beneath sea ice in areas of reduced thickness, increased coverage of melt ponds (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 number of cracks (high 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). Local features including snow-free or thin snow, hummocks and ridges commonly found on multi-year ice also provide habitat for ice algae (Lange et al., 2017).

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 (*medium confidence*) and these indirect impacts will possibly increase in the future (*low confidence*). 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 (Section 3.2.1.2.2, SM3.2.2) and decreased upwelling of nutrients into surface waters (Capotondi et al., 2012; Nummelin et al., 2016) may reduce 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, which require less nutrients, 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 favour 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).

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) (Sections 3.2.3.1.2, 5.3.6, SM3.2.6). Phytoplankton may have the capacity to compensate for ocean acidification under a range of temperatures and pH values (Hoppe et al., 2018).

Increased water temperatures (Section 3.2.1) and shifts in the spatial pattern and timing of the ice algal and phytoplankton blooms, have impacted the phenology, magnitude and duration of zooplankton production with associated changes in the zooplankton community composition (*medium confidence*). Negative effects of reductions in ice algae on zooplankton may be partially offset by predicted increases in water column phytoplankton production in the Bering Sea (Wang et al., 2015). Changes in sea ice coverage and thickness may alter the phenology, abundance and distribution of zooplankton in the future. Projected changes will initially have the most pronounced impact on sympagic amphipods, but will subsequently affect food web functioning and carbon dynamics of the pelagic system (Kohlbach et al., 2016).

At the more southern boundaries of the Arctic such as the southeastern Bering Sea, warm conditions have led to reduced production of large copepods and euphausiids (*medium confidence*) (Sigler et al., 2017; Kimmel et al., 2018). On more northern shelves, the increased open water period has 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) (*medium confidence*).

There are inconsistent findings concerning the future development of copepods in the Arctic. Coupled bio-physical model results suggest that sea ice loss will increase primary production and that will primarily be consumed pelagically by zooplankton grazers such as *Calanus hyperboreus*; increasing their abundances in the central Arctic (Kvile et al., 2018). Feng et al. (2018) concluded that *C. glacialis* should continue to benefit from a warmer Arctic Ocean. On the other hand, in the transition zone between Arctic and Atlantic water masses, *C. glacialis* may face increasing competition from the more boreal *C. finmarchicus* (Dalpadado et al., 2016). Renaud et al. (2018) found the lipid content of *Calanus* spp. was related to size and not species. This suggests that climate driven shifts in dominant *Calanus* species may, because of overlap in size spectrum and contrary to earlier assumptions, not negatively impact their consumers in the Barents Sea.

The effects of ocean acidification on Arctic zooplankton and pteropods (small pelagic molluscs) have been examined for only a few species and these studies reveal that the severity of effects is dependent on emission scenarios and the species sensitivity and adaptive capacity. The copepod *Calanus glacialis* exhibits stage-specific sensitivities to ocean acidification with some stages being relatively insensitive to decreases in pH and other stages exhibiting substantial reductions in scope for growth (Bailey et al., 2016; Thor et al., 2018). Although there is strong evidence that pteropods are sensitive to the effects of ocean acidification (Manno et al., 2017) recent studies indicate they may exhibit some ability to adapt (Peck et al., 2016; Peck et al., 2018). However, the metabolic costs of adaptation may be constraining, especially during periods of low food availability (Lischka and Riebesell, 2016).

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, 3.2.1.2) are impacting the benthic species composition (Birchenough et al., 2015). Other human-influenced activities, such as commercial bottom trawling and the introduction of non-native 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.

Rapid and extensive structural changes in the rocky-bottom communities of two Arctic fjords in the Svalbard Archipelago during the period 1980 to 2010 have been documented and linked to gradually increasing seawater temperature and decreasing sea ice cover (Kortsch et al., 2012; Kortsch et al., 2015). Also, there are indications 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 (Chapter 6, Table 6.1; 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 (Kortsch et al., 2012; Paar et al., 2016) (*medium confidence*) (See Section 5.3.3 and SM3.2.6 for further information on kelp).

The growth, early survival and production of commercially-important crab stocks in the Bering Sea are influenced by time-varying exposure to multiple interacting drivers including bottom temperature, larval advection, predation, competition, and fishing (Burgos et al., 2013; Long et al., 2015; Ryer et al., 2016). In Newfoundland and Labrador waters and on the western Scotian Shelf, snow crab (*Chionoecetes opilio*) productivity has declined (Mullowney et al., 2014; Zisserson and Cook, 2017). Contrary to this, snow crabs have expanded their distribution in the Barents Sea and commercial harvesting increased (Hansen, 2016; Lorentzen et al., 2018) (*high confidence*).

Bering sea crabs exhibit species-specific sensitivities to reduced pH (Long et al., 2016; Swiney et al., 2017; Long et al., 2019). However, current pH levels do not appear to have negatively impacted crab production in the Bering or Barents Seas (Mathis et al., 2015; Punt et al., 2015).

3.2.3.1.3 Fish

Since AR5, additional evidence shows climate induced physical and biogeochemical changes are impacting, and will continue to impact, the distribution and production of marine fish (*medium confidence*). 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 3.4; Figure 3.5), while data is severely limited in other Arctic shelf regions and the Central Arctic Ocean.

Higher temperature and changes in the quality and distribution of prey is already affecting marine fish (Wassmann et al., 2015; Dalpadado et al., 2016; Hunt et al., 2016; Section 3.2.3.1) (*high confidence* for detection*, medium confidence* for attribution). In the northern Barents Sea, Atlantic sector, higher temperatures (Section 3.2.1.2) have expanded suitable feeding areas for boreal/subarctic species (Box 3.4) and has 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 affected negatively 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). These changes may cause structural changes in food webs, with large piscivorous and semipelagic boreal fish species replacing small-bodied Arctic benthivores (Box 3.4; Fossheim et al., 2015; Frainer et al., 2017).

Time series on responses of anadromous fish (including salmon) in the high Arctic are 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 growth and winter survival (Jensen et al., 2017).

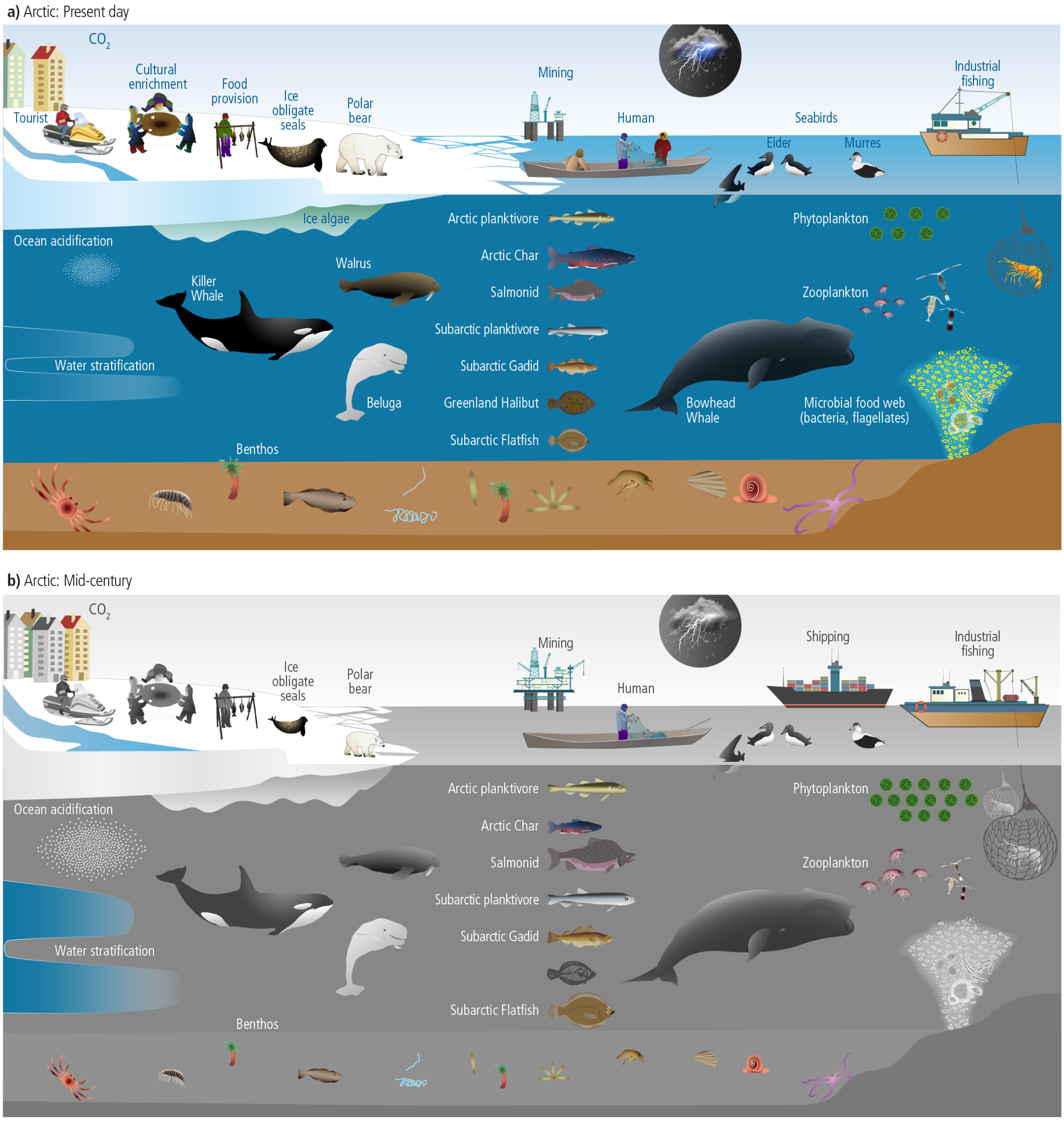


Figure 3.5: Schematic of Arctic system in present day (Panel a) and projected changes by mid-century (Panel b). Coloured elements in Panel b correspond to projected changes.

The scope for adaptation of marine fish to a changing ocean conditions is uncertain, but knowledge is informed by 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; Alabia et al., 2018) (*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 subarctic species that spawn in positive temperatures onto the high Arctic shelves (Landa et al., 2014). Projected increases in summer temperature 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 (Mueter et al., 2017; Joli et al., 2018). For example, the pelagic capelin (*Mallotus villosus*) are capable of entering the central Arctic Ocean, but may be restricted in winter by availability of suitable spawning areas and lack of antifreeze proteins (Hop and Gjøsæter, 2013; Christiansen, 2017).

Regional climate scenarios, derived from downscaled global climate scenarios, have been used to drive environmentally-linked fish population models (Hermann et al., 2016; Holsman et al., 2016; Ianelli et al., 2016; Hermann et al., 2019). Hermann et al. (2019) contrasted future production of copepods and euphausiids in the eastern Bering Sea under scenarios derived from projected downscaled high spatial and temporal resolution ocean habitats under RCP4.5 and 8.5. Consistent with AR5, these updated scenarios project future declines in the abundance of large copepods under RCP8.5, a result that has been show to negatively impact production of walleye pollock, Pacific cod (*Gadus microcephalus*) and arrowtooth flounder (*Atheresthes stomias*) (Sigler et al., 2017; Kimmel et al., 2018) (*medium confidence*). Hedger et al. (2013) predicts increases in Atlantic salmon abundance in northern Norway (river Alta around 70°N) with future warming (*low confidence*). Under end of century RCP8.5 projections, ocean acidification and higher ocean temperatures are expected to reduce production of Barents Sea cod (Stiasny et al., 2016; Koenigstein et al., 2018) (*low confidence*).

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; Laidre et al., 2015; Gall et al., 2017) (*high confidence*). These changes include altered ecological interactions as well as direct responses to habitat degradation induced especially via loss of sea ice. Population responses to warming have not all been linear, some have been particularly strong and abrupt due to environmental regime shifts, as seen in black-legged kittiwakes (*Rissa tridactyla*). A steep population decline in kittiwake colonies distributed throughout their breeding range coincided with an abrupt warming of sea-surface temperature in the 1990s, while their population dynamics did not seem to be affected during periods of more gradual warming (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; Hamilton et al., 2019) (*very high confidence*). However, some species display strong site fidelity that can be maladaptive in a changing climate and Arctic endemic marine mammals (all of which are ice-affiliated for breeding) in general have little scope to move northward in response to warming (Kovacs et al., 2012; Hamilton et al., 2015). 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 (deHart and Picco, 2015; Hamilton 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 strandings and pup mortality in years with little ice (Johnston et al., 2012c; 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; Choy et al., 2017) (*very high confidence*). Sea ice changes have increased the foraging effort of ringed seals (*Pusa hispida*) in the marginal ice zone north of Svalbard (Hamilton et al., 2015), also causing diet shifts (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, *Eschrichtius 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*) (Andersen et al., 2012; Hamilton et al., 2017; Escajeda et al., 2018) (*high confidence*). 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 particularly dangerous for young cubs (Durner et al., 2017; Pilfold et al., 2017; Lone et al., 2018). Cumulatively, changes in sea ice patterns are driving demographic changes in polar bears, including declines in some populations (Lunn et al., 2016; McCall et al., 2016), while others are stable or increasing (Voorhees et al., 2014; Aars et al., 2017). This is because protective management measures have been successful in allowing severely depleted populations to recover or because new food sources, such as carrion, 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 (Øigård et al., 2014; Breed et al., 2017; Smith et al., 2017a).

Several studies from different parts of the Arctic show evidence that changing temperatures impact seabirds diets (Dorresteijn et al., 2012; Divoky et al., 2015; Vihtakari et al., 2018), reproductive success and body condition (Gaston et al., 2012; Provencher et al., 2012; Gaston and Elliott, 2014) (*high confidence*). Recent studies also show that changes in sea surface temperature and sea ice dynamics have impacts on the distribution and abundance of seabird prey with cascading impacts on seabird community composition (Gall et al., 2017), nutritional stress, and decreased reproductive output (Dorresteijn et al., 2012; Divoky et al.; Kokubun et al., 2018) and survival (Renner et al., 2016; Hunt et al., 2018).

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 7 in Chapter 3), 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) (Section 5.2.2.5). 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) (SM3.2.6). This is due in part to the high abundance and circumpolar distribution of Antarctic krill, 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., 2017). 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; Constable et al., 2016; Hunt et al., 2016; Gutt et al., 2017), and the direct responses of biota to these changes (Constable et al., 2014) (summarised in Figure 3.6). These findings indicate that overlapping changes in key ocean and sea-ice habitat characteristics (temperature, sea-ice cover, iceberg scour, mixed layer depth, aragonite undersaturation; Sections 3.2.1, 3.2.2) will be important in determining future states of Southern Ocean ecosystems (Constable et al., 2014; Gutt et al., 2015) (*medium confidence*). However, there is a need to better characterize the nature and importance of indirect responses to physical change using models and observations. Important advances have also been made since AR5 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 projections from global climate models in ecological studies and in ecosystem models for the Southern Ocean (Cavanagh et al., 2017).

3.2.3.2.1 Plankton and pelagic 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). Arrigo et al. (2008) also report no overall trend in remotely-sensed column-integrated primary production south of 50°S from 1998 to 2006. At a regional scale, local-scale forcings (e.g., retreating glaciers, topographically-steered circulation and sea ice duration) and associated changes in stratification are key determinants of phytoplankton bloom dynamics at coastal stations on the West Antarctic Peninsula (Venables et al., 2013; Schofield et al., 2017; Kim et al., 2018; Schofield et al., 2018) (*medium confidence*). For example, a shallowing trend in mixed layer depth in the southern part of the Peninsula (as opposed to no trend in the north) associated with changes in sea-ice duration over a 24 year period (from 1993 to 2017) has been linked to enhanced phytoplankton productivity (Schofield et al., 2018).

The phenology of Southern Ocean phytoplankton blooms in this region may also be shifting to earlier in the growth season (Arrigo et al., 2017a). However, the effect of climate change on Southern Ocean pelagic 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.

Recent studies on the ecological effects of acidification in coastal waters near the Antarctic continent indicate a detrimental effect of acidification on primary production and changes to the structure and function of microbial communities (Hancock et al., 2017; Deppeler et al., 2018; Westwood et al., 2018) (*medium confidence*). Trimborn et al. (2017) report that Southern Ocean diatoms are more sensitive to ocean acidification and changes in irradiance than the prymnesiophyte *Phaeocystis antarctica*, which may have implications for biogeochemical cycling because diatoms and prymnesiophytes are generally considered key drivers of these cycles. Both laboratory manipulations and *in situ* experiments indicate that sea-ice algae are tolerant to acidification (McMinn, 2017) (*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, from Leung et al. (2015). There is *low confidence* in predicted zonal changes in phytoplankton biomass due to *low confidence* regarding future changes in iron supply in the Southern Ocean (Hutchins and Boyd, 2016). 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 mean underwater irradiance More iron supply	Shallowing of the summertime mixed layer depth Change in iron supply mechanism
50°S–65°S	↓	Lower mean underwater irradiance	Deeper summertime mixed layer depth Decreased summertime incident radiation (increased cloud fraction)
S of 65°S	↑	More iron supply Higher mean underwater irradiance Temperature	Melting of sea–ice Warming ocean

Previously reported declines in Antarctic krill abundance in the South Atlantic sector (Atkinson et al., 2004) cited in WGII AR5 (Larsen et al., 2014) may not represent a long-term, climate-driven, regional-scale decline (Fielding et al., 2014; Kinzey et al., 2015; Steinberg et al., 2015; Cox et al., 2018) (*medium confidence*) but could reflect a sudden, discontinuous change following an episodic period of anomalous peak abundance for this species (Loeb and Santora, 2015) (*low confidence*). Recent analyses have not detected trends in long-term krill abundance in the South Atlantic sector in acoustic surveys (Fielding et al., 2014; Kinzey et al., 2015), net-based surveys (Steinberg et al., 2015) or re-analysis of historical data (Cox et al., 2018). Nevertheless, the spatial distribution and size composition of Antarctic krill may already have changed in association with change in the sea ice environment (Atkinson et al., 2019) (*medium confidence*) and may result in different regional trends in numerical krill abundance (Cox et al., 2018; Atkinson et al., 2019) (*medium 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 growth and recruitment (Melbourne-Thomas et al., 2016; Piñones and Fedorov, 2016; Meyer et al., 2017; Murphy et al., 2017; Klein et al., 2018). 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 greatest projected reductions in krill due to the effects of warming and ocean acidification 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 main 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 RCP8.5 (Klein et al., 2018). 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 (6, 15, and 41% increases in phytoplankton production with equivalent percentage decreases in ice algal production) 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). 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). Pteropods are vulnerable to the effects of acidification, and new evidence indicates that eggs released at high CO_2 concentrations lack resilience to ocean acidification in the Scotia Sea region (Manno et al., 2016) (*medium confidence*).

3.2.3.2.2 Benthic communities

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). Bentho-pelagic coupling and vertical energy flux will also influence Southern Ocean ecosystem responses to climate change (Jansen et al., 2017). Benthic 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 the 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*) (Table 6.1). Griffiths et al. (2017a) modelled distribution changes for 963 benthic invertebrate species in the Southern Ocean 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. Predicted reductions in the number of species are most pronounced for the West Antarctic Peninsula and the Scotia Sea region (Griffiths et al., 2017a).

3.2.3.2.3 Fish

Many Antarctic fish have a narrow thermal tolerance as a result of physiological adaptations to cold water (Pörtner et al., 2014; Mintenbeck, 2017), which makes them vulnerable to the effects of increasing temperatures (Mueller et al., 2012; Beers and Jayasundara, 2015). Increasing water temperatures may displace icefish (family *Channichthyidae*) in marginal habitats (e.g. shallow regions around subantarctic islands) as they lack haemoglobin and are unable to adjust blood parameters to an increasing oxygen demand (Mintenbeck et al., 2012) (*low confidence*). Future warming may also reduce the planktonic duration and increase egg and larval mortality for fish species, which is predicted to affect dispersal patterns, with implications for population connectivity and the ability of fish species to adapt to ongoing environmental change (Young et al., 2018). The Antarctic silverfish (*Pleuragramma antarctica*) 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. Species distribution models for *Electrona antarctica*, a dominant myctophid species in the Southern Ocean, project habitat loss for this species under RCP4.5 (6.2 ± 6.0% loss) and RCP8.5 (13.1 ± 10.2% loss) by 2090, associated with increased sea surface temperature (Freer et al., 2018). There have been no observed 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*), but recruitment is inversely correlated with sea surface temperature for Patagonian toothfish at South Georgia (Belchier and Collins, 2008). 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

Since AR5, there has been an increasing body of evidence of climate-induced changes in populations of some Antarctic higher predators such as seabirds and marine mammals. These changes vary between different regions of the Southern Ocean and reflect differences in key drivers (Bost et al., 2009; Gutt et al., 2015; Constable et al., 2016; Hunt et al., 2016; Gutt et al., 2017), particularly sea-ice extent and food availability (*high confidence)* across regions (Sections 3.2.1.1.1, 5.2.3.1, 5.2.3.2, 5.2.4). The predictability of foraging grounds and ice cover are associated with variations in climate (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 higher 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, body 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 (Braithwaite et al., 2015a; Whitehead et al., 2015; Seyboth et al., 2016; Hinke et al., 2017b) (*high confidence*).

Trends of populations of Antarctic penguins affected by climate change 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*). Yet population shifts in Adélie penguins (Youngflesh et al., 2017) may have resulted from strong interannual environmental variability in good and bad years for prey and breeding habitat rather than climate change (*low confidence*). 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 (seasonally and in specific regions), increases in sea surface temperatures, extreme events (snow storms) and wind regime shifts 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*) (Descamps et al., 2015; Jenouvrier et al., 2015; Pardo et al., 2017) (*low confidence)*. Poleward population shifts with increased intensity and frequency of westerly winds affect functional traits, demographic rates, foraging range, rates of travel and flight speeds of flying birds (Weimerskirch et al., 2012; Jenouvrier et al., 2018) but also increase overlap with fisheries activities thus increasing the risk of bycatch and the need for mitigation measures (Krüger et al., 2018) (*medium confidence)*.

Changes in local and regional-scale oceanographic features (Section 3.2.1.2) together with bathymetry control prey aggregation and distribution, and affect the ecological responses and biological traits of higher predators (particularly 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; Hinke et al., 2017a; Pardo et al., 2017) (*medium confidence*) and *likely* explain most of the 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 (*Ommatophoca rossi)*, crabeater seals (*Lobodon carcinophaga)*, leopard seals (*Hydrurga leptonyx)* and Weddell seals (*Leptonychotes weddellii)*) as a reference point for understanding climate change impacts on these species (Southwell et al., 2012; Bester et al., 2017), although some regional population estimates for pack ice seals are available (Gurarie et al., 2017) and references therein). 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*). There have been predictions of negative future impacts of climate change on krill and all whale species, although the magnitude of impacts differs among populations (Tulloch et al., 2019) as for other higher predators (Section 5.2.3). Pacific blue (Tulloch et al., 2019) (*Balaenoptera musculus*), fin (*B. physalus*) and southern right whales (*Eubalaena australis*) are the most at risk but humpback whales (*Megaptera novaeangliae*) are also at risk, as consequence of reduced prey and increasing interspecific competition. Importantly, climate-related risks for whale populations are a product of environmental conditions and connectivity between whale foraging grounds (Southern Ocean) and breeding grounds (lower latitudes) (Section 5.2.3.1).

3.2.3.2.5 Pelagic foodwebs and ecosystem structure

This section assesses the impacts of ocean and sea ice changes on pelagic foodwebs and ecosystem structure. The ecological impacts of loss of ice shelves and retreat of coastal glaciers around Antarctica are assessed in Section 3.3.3.4. 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; Constable et al., 2016; Constable et al., 2017; McCormack et al., 2017) (*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 declines in krill predator populations, particularly penguins, 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) (Constable et al., 2016; Constable et al., 2017) (Sections 3.2.4.1.2, 3.5.3.2.2).

In summary, advances in knowledge regarding the impacts of climate change on Antarctic marine ecosystems since AR5 are consistent with the impacts described in Larsen et al. (2014) (also summarized in Figure 3.6). These advances include further descriptions of local-scale, climate-related influences (sea ice and stratification) on primary productivity, particularly in the West Antarctic Peninsula region (Section 3.2.3.2.1) (*medium confidence*). At the circumpolar scale, primary production is projected to increase in regions south of 65°S over the period from now to 2100 under RCP8.5 (Leung et al., 2015) (*low confidence*). However, ocean acidification may have a detrimental effect on coastal phytoplankton communities around the Antarctic continent (Section 3.2.3.2.1) (*medium confidence*). Increased information is also available regarding climate-driven changes in Antarctic krill populations in the south Atlantic, including the observed southward shift in the spatial distribution of krill in this region (Atkinson et al., 2019) (*medium confidence*) but evidence of a long-term trend in overall abundance in the region is equivocal (Section 3.2.3.2.1). Further habitat contraction for Antarctic krill is predicted in the future (*medium confidence*) (references detailed in Section 3.2.3.2.1). Under high emissions scenarios the majority of Antarctic seafloor species are projected to be negatively impacted by the end of the century (Griffiths et al., 2017a) (*low confidence*). Observed changes in the geography of ice-associated habitats (sea ice, ice shelves and polynyas) have both positive and negative effects on sea birds and marine mammals, and will interact with ice-dependent changes in Antarctic krill populations to compound the impacts on krill-dependent predators (Klein et al., 2018) (Sections 3.2.3.2.1, 3.2.3.2.4) (*medium confidence*).

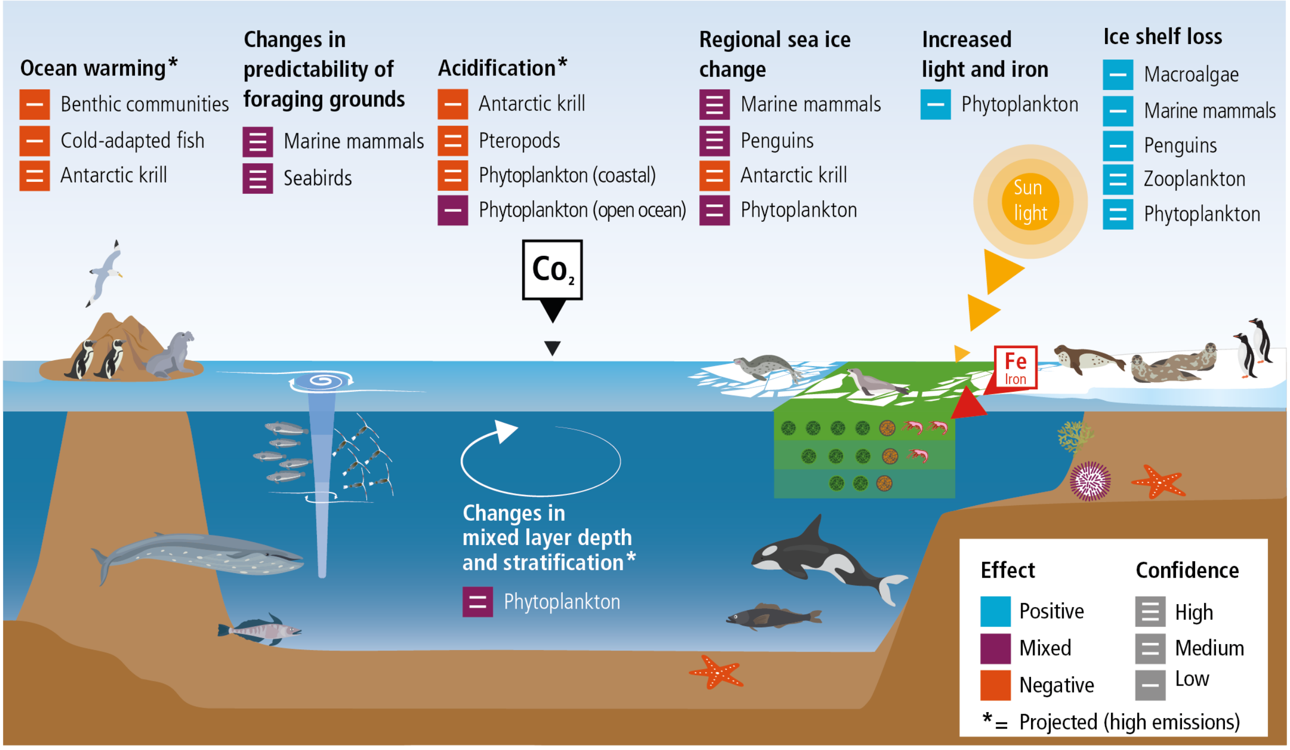


Figure 3.6: Schematic summary of key drivers that are causing or are projected to cause direct effects on Southern Ocean marine ecosystems. Effects presented here are described in the main text (Sections 3.2.3.2, 3.3.3.4), with associated confidence levels and citations. Projected changes (indicated by an asterisk) are for high emissions scenarios. The cross-sectional view of the Southern Ocean ecosystem shows the association of key functional groups (marine mammals, birds, fish, zooplankton, phytoplankton and benthic assemblages) with Southern Ocean habitats. The configuration of the Southern Ocean foodweb is described in SM3.2.6.

3.2.4 Impacts on Social-Ecological Systems

3.2.4.1 Fisheries

3.2.4.1.1 Arctic

Arctic fisheries are important economically and societally. Large commercial fisheries exist off the coasts of Greenland and in the Barents and Bering Seas (Holsman et al., 2018; Peck and Pinnegar, 2018). First-wholesale value for commercial harvest of all species in 2017 in the Eastern Bering Sea was $2.68 billion and for the Barents Sea around US$1 billion to Norwegian fishers alone. The target species for these commercial fisheries include gadoids, flatfish, herring, red fish (*Sebastes* sp.), salmonids, and capelin. Fisheries in other Arctic regions are relatively small-scale, locally operated, and target a limited number of species (Reist, 2018). Still, these fisheries are of considerable cultural, economic, and subsistence importance to local communities (Section 3.5.2.1).

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*). Past performance 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 a thousand years (Barrett et al., 2011), reflecting the resilience of the northern Norwegian cod fisheries to historic climate variability (Eide, 2017). Also, 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 (Lam et al., 2016; Eide, 2017; Haug et al., 2017; Peck and Pinnegar, 2018) (Section 3.2.3.1.3, Box 3.4) (*medium confidence*).

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 of 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 Arctic shelf regions (Section 3.2.3.1) (*medium confidence*).

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) (Table 3.4). Further, past experience suggests that barriers to diversification may limit the portfolio of viable target fisheries available to small-scale fisheries (Ward et al., 2017) (*low confidence*).

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 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 mean annual wholesale value of the Antarctic krill fishery was US$69.5 million per year for the period from 2011-2015, and US$206.7 million per year for toothfish fisheries (combined) over the same period (CCAMLR, 2016b). 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 predators (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) (*low 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). Available evidence regarding future changes to Antarctic krill populations (Section 3.2.3.2.1) 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 and socio-economic drivers.

There is limited understanding of 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 to an increase in accessibility (Dawson et al., 2014; Johnston et al., 2017) (*high confidence*). While not exclusively ‘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 (Johnston 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, more purpose-built polar cruise vessels are being constructed, and private pleasure craft are appearing in the Arctic more frequently (Lasserre and Têtu, 2015; Johnston et al., 2017; Dawson et al., 2018). In Antarctica, almost 37,000 (predominantly shipborne) tourists visited in 2016/17, with 51,707 during 2017/18; there were 6,700 tourists in 1992/93 (the first year of record) (ATCM, 2018). 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).

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). There is *high confidence* that demand for Arctic cruise tourism will continue to grow over the coming decade (Johnston et al., 2017). The anticipated implications of future climate change have become a driver for polar tourism. A niche market known as 'last chance tourism' has emerged 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 within communities, 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). There is *high confidence* that biodiversity supported by ice-free areas, particularly those on the Antarctic Peninsula, are vulnerable to the introduction of terrestrial alien species via tourists and scientists (Chown et al., 2012; Huiskes et al., 2014; Hughes et al., 2015; Duffy et al., 2017; Lee et al., 2017a) (Box 3.3) as well as to the direct impacts of humans (Pertierra et al., 2017). The tourism sector relies on a set of regulations that apply to all types of maritime shipping, yet cruise ships intentionally travel off regular shipping corridors and serve a very different purpose than other vessel types, so there is a need for region-specific governance regimes, specialized infrastructure, and focused policy attention (Dawson et al., 2014; Pashkevich et al., 2015; Pizzolato et al., 2016; Johnston 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. At the same time, 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, the Arctic Bridge and the 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 *high confidence* that shipping activity during the Arctic summer increased over the past two decades in regions for which there is information, concurrent with reductions in Arctic sea ice extent and the shift to predominantly seasonal ice cover (Pizzolato et al., 2014; Eguíluz et al., 2016; Pizzolato et al., 2016). Long term datasets over the pan-Arctic are incomplete, but the distance travelled by ships in Arctic Canada nearly tripled between 1990 and 2015 (from ~365,000 km to ~920,000 km) (Dawson et al., 2018). Other non-environmental factors which influence Arctic shipping are 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; Johnston et al., 2017). Current impacts associated with the observed increase in Arctic shipping include a higher rate of reported accidents per km travelled compared to southern waters (CCA, 2016), increases in vessel noise propagation (Halliday et al., 2017) and air pollution (Marelle et al., 2016). Disruptions to cultural and subsistence hunting activities from increased shipping (Huntington et al., 2015; Olsen et al., 2019) compound climate-related impacts to people (Sections 3.4.3.3.2, 3.4.3.3.3).

It is projected that shipping activity will continue to rise across the Arctic as northern routes become increasingly accessible (Stephenson et al., 2011; Stephenson et al., 2013; Barnhart et al., 2015; Melia et al., 2016), although mitigating economic and operational factors remain uncertain and could influence future traffic volume (Zhang et al., 2016). The Northern Sea Route is expected to be more viable than other routes because of infrastructure already in place (Milaković et al., 2018); favourable summer ice conditions in recent years have reduced transit times (Aksenov et al., 2017). In comparison, the Northwest Passage and

Arctic Bridge presently have limited port and marine transportation infrastructure, incomplete soundings and hydrographic charting, challenging sea ice conditions, and limited search and rescue capacity; these compound the risks from shipping activity (Stephenson et al., 2013; Johnston et al., 2017; Andrews et al., 2018).

Future shipping impacts will be regionally diverse considering the unique geographies, sea ice dynamics, infrastructure and service availability, and regulatory regimes that exist across different Arctic nations. Considerations include socio-economic and political implications related to safety (marine and local accidents), security (trafficking, terrorism, local issues), 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, 2015a; Halliday et al., 2017; Hauser et al., 2018). Black carbon emissions from shipping activity within the Arctic are projected to increase (Arctic Council, 2017) and are more easily deposited at the surface in the region compared with emissions from lower latitudes (Sand et al., 2013). Commercial shipping mainly uses heavy fuel oil, with associated emissions of sulphur, nitrogen, metals, hydrocarbons, organic compounds, black carbon and fly ash to the atmosphere during combustion (Turner et al., 2017a). Mitigation approaches include banning heavy fuel oil as already implemented in Antarctica and the waters around Svalbard, and the use of new technology like scrubbers.

The predominant shipborne activities in Antarctica are fishing, logistic support to land-based stations, and marine research vessels operating for both non-governmental and governmental sectors. Uncertainty in future Antarctic sea ice conditions (Section 3.2.2.1) pose challenges to considering potential impacts on these activities (Chown, 2017).

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

3.3.1 Ice Sheet Changes

Changes in ice sheet mass have been derived repeatedly over the satellite era using complementary methods based on time series of satellite altimetry to measure volume change, ice-flux measurements combined with modelled surface mass balance to calculate mass inputs and outputs, and satellite gravimetry to measure regional mass change. Ice sheet changes over earlier periods have also been reconstructed from firn/ice-core and geological evidence (SM3.3.1).

3.3.1.1 Antarctic Ice Sheet Mass Change

It is *virtually certain* that the Antarctic Peninsula (AP) and West Antarctic Ice Sheet (WAIS) combined have cumulatively lost mass since widespread measurements began in 1992, and that the rate of loss has increased since around the year 2006 and continued post-AR5 (Martín-Español et al., 2016; Zwally et al., 2017; Bamber et al., 2018; Gardner et al., 2018; The IMBIE Team, 2018; Rignot et al., 2019), extending and reinforcing previous findings (IPCC, 2013) (Figure 3.7, Table 3.3, SM3.3.1.1). From *medium evidence*, there is *high agreement* in the sign and *medium agreement* in the magnitude of both WAIS and AP mass change between the complementary satellite methods (Mémin et al., 2015; The IMBIE Team, 2018).

Table 3.3: Mass balance (Gt yr^{-1}) of the West Antarctic Ice Sheet (WAIS), Antarctic Peninsula (AP), East Antarctic Ice Sheet (EAIS), the combined Antarctic Ice Sheets (AIS) and the Greenland Ice Sheet (GIS) and the total sea level contribution (mm yr^{-1}).

Ice sheet	1992–1996	1997–2001	2002-2006	2007–2011	2012–2016
WAIS and AP (Bamber et al., 2018)	-55 ±30	-53 ±30	−77 ±17	−197 ±11	−172 ±27
WAIS and AP (The IMBIE Team, 2018)	-60 ±32	-44 ±31	-85 ±31	-183 ±32	-192 ±31
WAIS only (The IMBIE Team, 2018)	-53 ±29	-41 ±28	-65 ±27	-148 ±27	-159 ±26
EAIS (Bamber et al., 2018)	28 ±76	−50 ±76	52 ±37	80 ±17	−19 ±20
EAIS (The IMBIE Team, 2018)	11±58	8 ±56	12 ±43	23 ±38	-28 ±30
GIS (Bamber et al., 2018)	31± 83	−47 ± 81	−206 ±28	−320 ±10	−247 ±15

	1992-2006		2007-2016	
WAIS and AP (Bamber et al., 2018; The IMBIE Team, 2018)	-56 ±20		-185 ±17	
	1992-2016			
EAIS (Bamber et al., 2018)	+18 ±52			
EAIS (The IMBIE Team, 2018)	+15 ±41			
	1992-2001	**2002-2011**	**2006-2015**	**2012-2016**
AIS (Bamber et al., 2018; The IMBIE Team, 2018)	-51 ±73	-82 ±27	-155 ±19	-199 ±26
GIS (Bamber et al., 2018)	-8 ±82	-263 ±21	-278 ±11	-247 ±15
Total sea level contribution (mm yr^{-1})	0.16 ± 0.3	0.96 ±0.1	1.20 ±0.1	1.24 ±0.1

WAIS mass loss and recent increases in loss were concentrated in the Amundsen Sea Embayment (ASE) (*high confidence*) with increases particularly in the late 2000s (Mouginot et al., 2014), accounting for most of the –112 ± 10 Gt yr^{-1} WAIS loss from 2003–2013 (Martín-Español et al., 2016). The ice-sheet margins of nearby Getz Ice Shelf also lost mass rapidly (–67 ± 27 Gt yr^{-1}, 2008–2015) (Gardner et al., 2018). This region also experienced losses during previous warm periods (Cross-Chapter Box 8 in Chapter 3).

On the AP, the Bellingshausen Sea ice sheet margin shifted from close to mass balance in the 2000s to rapid loss from 2009 (-56 ±8 Gt yr^{-1} from 2010-2014) (*high confidence*) (Helm et al., 2014; McMillan et al., 2014b; Wouters et al., 2015; Hogg et al., 2017). This shift accompanied ongoing mass loss (*high confidence*) from the smaller north-eastern AP glaciers that fed the former Prince Gustav, Larsen A and B ice shelves, though now at a lower rate than immediately following shelf collapse in 1995 and 2002 (Seehaus et al., 2015; Wuite et al., 2015; Rott et al., 2018). Of 860 marine-terminating AP glaciers, 90% retreated from their 1940s positions (Cook et al., 2014), established in the early to mid-Holocene (Ó Cofaigh et al., 2014) (*medium confidence*). Early 21st century combined AP glacier (Fieber et al., 2018) and ice sheet loss was around –30 Gt yr^{-1} (Table 3.3).

The East Antarctic Ice Sheet (EAIS, covering 85% of the AIS) has remained close to balance, with large interannual variability and no clear mass trend over the satellite record (*medium confidence*) (Table 3.3, Figure 3.7, SM3.3.1.2), and relatively large observation uncertainties (SM3.3.1) (Velicogna et al., 2014; Martin-Español et al., 2017; Bamber et al., 2018). Surface mass balance (SMB) trends are particularly ambiguous, leading to disagreement between one altimetry and one flux-based estimate of +136 ±43 Gt a^{-1} (spanning 1992-2008) (Zwally et al., 2017), and -41 ±8 Gt a^{-1} (1979-2017) (Rignot et al., 2019), respectively. Both differ from the multi-method averages reported here (Table 3.3).

EAIS mass gains on the Siple Coast and Dronning Maud Land (e.g., +63 ± 6 Gt yr^{-1} from 2003–2013 (Velicogna et al., 2014)) contrast with Wilkes Land losses e.g., from –17 ± 4 Gt yr^{-1} from the Totten Glacier area, 2003–2013 (Velicogna et al., 2014) that drain a large area of deeply-grounded EAIS with potential for multi-metre sea level contributions (Zwally et al., 2017; Rignot et al., 2019). Limited palaeo ice sheet evidence suggests that this area has previously lost substantial mass in previous interglacials (*medium confidence*) (Aitken et al., 2016; Wilson et al., 2018).

Overall, 2012-2016 AIS mass losses were *extremely likely* greater than those from 2002-2011 and *likely* greater than from 1992-2001, and it is *extremely likely* that the negative 2012-2016 AIS mass balance was dominated by losses from WAIS (Table 3.3).

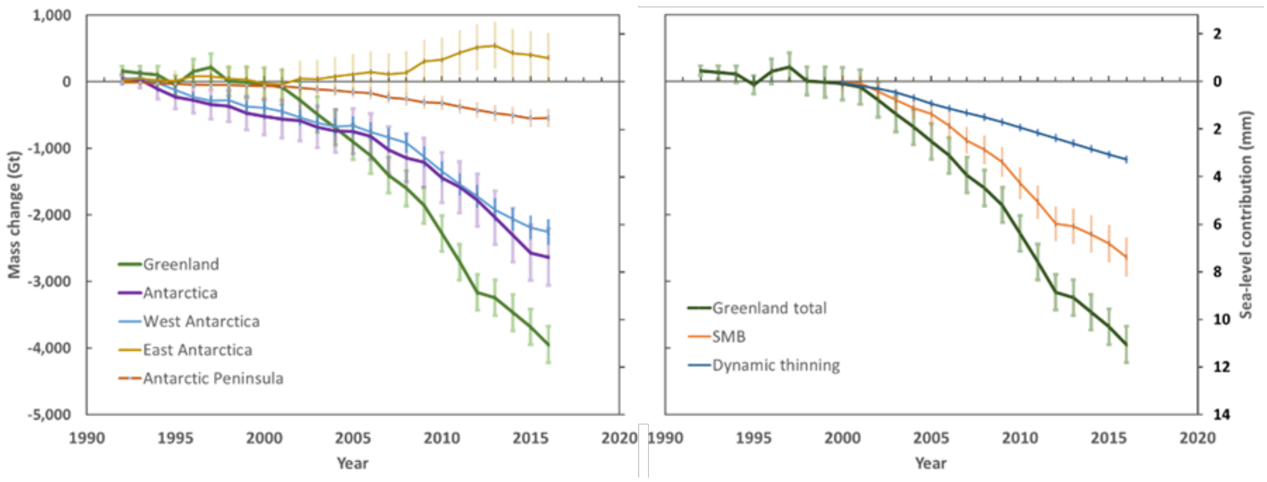


Figure 3.7: (a) Cumulative Ice Sheet mass change, 1992 to 2016, (after Bamber et al., 2018; The IMBIE Team, 2018). (b) Greenland Ice Sheet mass change components from surface mass balance (orange) and dynamic thinning (blue) from 2000-2016, (after Van den Broeke et al., 2016; King et al., 2018). Uncertainties are 1 standard deviation.

3.3.1.2 *Components of Antarctic Ice Sheet Mass Change*

AIS mass changes are dominated by changes in snowfall and glacier flow. The WAIS and AP loss trends in recent decades are dominated by glacier flow acceleration (also known as dynamic thinning) (*very high confidence*) (Figure SM3.8). Dynamic thinning losses were –112 ± 12 Gt yr^{-1} for 2003–2013, largely from the ASE (Figure SM3.8) (Martín-Español et al., 2016), which contributed –102 ± 10 Gt yr^{-1} from 2003–2011 (Sutterley et al., 2014). Total ASE ice discharge increased by 77% since the 1970s (Mouginot et al., 2014), primarily from 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). Dynamic thinning 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., 2018). Glacier acceleration of up to 25% also affected the Getz Ice Shelf margin from 2007-2014 (Chuter et al., 2017).

Reduction or loss of ice-shelf buttressing has dominated AIS dynamic thinning (*high confidence*). Ice shelves buttress 90% of AIS outflow (Depoorter et al., 2013; Rignot et al., 2014; Fürst et al., 2016; Reese et al., 2018), and ice-shelf thinning increased in WAIS by 70% in the decade to 2012, averaged 8% thickness loss from 1994–2012 in the ASE (Paolo et al., 2015), and explains the post-2009 onset of rapid dynamic thinning on the southern-AP Bellingshausen Sea coast (Wouters et al., 2015; Hogg et al., 2017; Martin-Español et al., 2017) (Figure SM3.8). Grounding-line retreat, an indicator of thinning, has been observed with *high confidence* (Rignot et al., 2014; Christie et al., 2016; Hogg et al., 2017; Konrad et al., 2018; Roberts et al., 2018). From 2010-2016, 22%, 3% and 10% of grounding lines in WAIS, EAIS and the AP respectively retreated at rates faster than 25 m yr^{-1} (the average pace since the Last Glacial Maximum; Konrad et al., 2018), with highest rates along the Amundsen and Bellingshausen Sea coasts, and around Totten Glacier, Wilkes Land, EAIS (Konrad et al., 2018), where dynamic thinning has occurred at least since 1979 (Roberts et al., 2018; Rignot et al., 2019). Ice-shelf collapse has driven dynamic thinning in the northern AP over recent decades (*high confidence*) (Seehaus et al., 2015; Wuite et al., 2015; Friedl et al., 2018; Rott et al., 2018).

ASE ice-shelf basal melting, grounding-line retreat and dynamic thinning have varied with ocean forcing (*medium confidence*) (Dutrieux et al., 2014; Paolo et al., 2015; Christianson et al., 2016; Jenkins et al., 2018) but this variability is superimposed on sustained mass losses compatible with the onset of marine ice sheet instability for several major glaciers (*medium confidence*) (Favier et al., 2014; Joughin et al., 2014; Mouginot et al., 2014; Rignot et al., 2014; Christianson et al., 2016). Whether unstable WAIS retreat has begun or is imminent remains a critical uncertainty (Cross-Chapter Box 8 in Chapter 3).

Mass gains due to increased snowfall have somewhat offset dynamic-thinning losses (*high confidence*). On the AP, snowfall began to increase in the 1930s, accelerated in the 1990s (Thomas et al., 2015; Goodwin et al., 2016), and now offsets sea-level rise by 6.2 ± 1.7 mm per century (Medley and Thomas, 2018). EAIS and WAIS snowfall increases offset 20th century sea-level rise by 7.7 ± 4.0 mm and 2.8 ± 1.7 mm respectively (Medley and Thomas, 2018) (*medium confidence*). AIS snowfall increased by +4 ± 1 then +14 ± 1 Gt per decade over the 19th and 20th centuries, of which EAIS contributed 10% (Thomas et al., 2017b). Longer records suggest either an AIS snowfall decrease over the last 1000 years (Thomas et al., 2017a) or a statistically negligible change over the last 800 years (*low confidence*) (Frezzotti et al., 2013).

Mass balance contributions from ice-sheet basal melting were not described in AR5 (IPCC, 2013) and the sensitivity of the AIS subglacial hydrological system to climate change is poorly understood. Around half of the AIS bed melts (Siegert et al., 2017), producing ~65 Gt yr^{-1} of water (Pattyn, 2010) (*low confidence*), some of which refreezes (Bell, 2008) and some accumulates in subglacial lakes with a total volume of tens of thousands of cubic kilometres (Popov and Masolov, 2007; Lipenkov et al., 2016; Siegert, 2017). This system contributes fresh water and nutrients to 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), and lubricates glacier sliding (e.g., Dow et al., 2018b). Changes in the ice sheet thickness can redistribute subglacial water, affecting drainage pathways and ice flow (Fricker et al., 2016), but hydrological observations are very scarce.

3.3.1.3 Greenland Ice Sheet Mass Change

The Greenland Ice Sheet (GIS) experienced a marked shift to strongly negative mass balance between the early 1990s and mid–2000s (*very high confidence*) (Shepherd et al., 2012; Schrama et al., 2014; Velicogna et al., 2014; Van den Broeke et al., 2016; Bamber et al., 2018; King et al., 2018; Sandberg Sørensen et al., 2018; WCRP, 2018). It is *extremely likely* that the 2002-2011 and 2012-2016 ice losses were greater than in the 1992-2001 period (Bamber et al., 2018) (Table 3.3, Figure 3.7, SM3.3.1.3). GIS mass balance is characterised by large interannual variability (e.g., van den Broeke et al., 2017) but from 2005-2016 GIS was the largest terrestrial contributor to global sea level rise (WCRP, 2018).

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 northwest and southeast coasts, and more locally in the southwest and on the large west-coast Jakobshavn Glacier, though intensifying and spreading to the remainder of the coastal ice sheet in the latest period (Kjeldsen et al., 2015). Palaeo evidence also suggests that the GIS has contributed substantially to sea level rise during past warm intervals (Cross-Chapter Box 8 in Chapter 3).

3.3.1.4 Components of Greenland Ice Sheet Mass Change

Ongoing GIS mass loss over recent years has resulted from a combined increase in dynamic thinning and a decrease in SMB. Of these, reduced SMB due to an increase in surface melting and runoff recently came to dominate (*high confidence*) (Andersen et al., 2015; Fettweis et al., 2017; van den Broeke et al., 2017; King et al., 2018), accounting for 42% of losses for 2000–2005, 64% for 2005–2009 and 68% for 2009–2012 (Enderlin et al., 2014) (Figure 3.7).

The GIS was close to balance in the early years of the 1990s (Hanna et al., 2013; Khan et al., 2015), the interior above 2000 m altitude gained mass from 1961–1990 (Colgan et al., 2015) and both coastal and ice-sheet sites experienced an increasing precipitation trend from 1890 to 2012 and 1890 to 2000 respectively (Mernild et al., 2015), but since the early 1990s multiple observations and modelling studies show strong warming and an increase in runoff (*very high confidence*). High-altitude GIS sites NEEM and Summit warmed by, respectively, 2.7 ± 0.33°C over the past 30 years (Orsi et al., 2017) and by 2.7 ± 0.3°C from 1982–2011 (McGrath et al., 2013), while satellite thermometry showed statistically significant widespread surface warming over northern GIS from 2000-2012 (Hall et al., 2013). The post–1990s period experienced the warmest GIS near-surface summer air temperatures of 1840–2010 (+1.1˚C) (statistically highly significant) (Box, 2013), and ice core analysis found the 2000-2010 decade to be the warmest for around 2000 years (Vinther et al., 2009; Masson-Delmotte et al., 2012), and possibly around 7000 years (Lecavalier et al., 2017). This significant summer warming since the early 1990s increased GIS melt-event duration (Mernild et al., 2017) and intensity to levels exceptional over at least 350 years (Trusel et al., 2018), and melt frequency to levels unprecedented for at least 470 years (Graeter et al., 2018). GIS melt intensity for 1994-

2013 was two-to-fivefold the pre-industrial intensity (*medium confidence*) (Trusel et al., 2018). In response, GIS meltwater production and runoff increased (Hanna et al., 2012; Box, 2013; Fettweis et al., 2013; Tedstone et al., 2015; Van den Broeke et al., 2016; Fettweis et al., 2017), resulting in 1994-2013 runoff being 33% higher the 20th century mean and 50% higher than the 18th century (Trusel et al., 2018), and 80% higher in a western-GIS marginal river catchment in 2003-2014 relative to 1976-2002 (Ahlstrom et al., 2017).

Only around half of the 1960-2014 surface melt ran off, most of the rest being retained in firn and snow (Steger et al., 2017), particularly in recently-observed firn aquifers in south and west Greenland (Humphrey et al., 2012; Forster et al., 2013; Kuipers Munneke et al., 2014; Poinar et al., 2017) that cover up to 5% of GIS (Miège et al., 2016; Steger et al., 2017) and stored around one fifth of the meltwater increase since the late 1990s (Noël et al., 2017) (*medium confidence*). While potential aquifer storage is equivalent to about a quarter of annual GIS melt production (Koenig et al., 2014; Van den Broeke et al., 2016) and aquifers have spread to higher altitudes (Steger et al., 2017), their potential to buffer runoff has been reduced by firn densification (Polashenski et al., 2014), diversion of water to the bed via crevasses (Poinar et al., 2017), and the formation of ice layers that prevent drainage and promote surface ponding on the firn (Charalampidis et al., 2016) (*high confidence*). Such ponding lowers the firn albedo, promoting further melting (*high confidence*) (e.g., Charalampidis et al., 2015), but the extent of bare ice is a fivefold stronger control on melt (Ryan et al., 2019). Bare ice produced ~78% of runoff from 1960-2014, and its extent is expected to increase non-linearly as snow cover retreats to higher, flatter areas of ice sheet (Steger et al., 2017). This extent is not well reproduced in climate models, however, with biases of -6% to +13% (Ryan et al., 2019).

The remaining ~40% of non-SMB GIS mass loss from 1991 to 2015 has resulted from increased ice discharge due to dynamic thinning (*high confidence*) (Enderlin et al., 2014; Van den Broeke et al., 2016; King et al., 2018) (Figure 3.7). From 2000 to 2016, dynamic thinning of 89% of GIS outlet glaciers accounted for –682 ± 31 Gt mass change, of which 92% came from the northwest and southeast GIS (King et al., 2018). Half came from only four glaciers (Jakobshavn Isbræ, Kangerdlugssuaq, Koge Bugt, and Ikertivaq South) (Enderlin et al., 2014). Glacier thinning has decreased glacier discharge, however, reducing the dynamic contribution to GIS mass loss (e.g., from 58% from 2000 to 2005 to 32% between 2009 and 2012; Enderlin et al., 2014). Furthermore, there is now *high confidence* that for most of the GIS, increased surface melt has not led to sustained increases in glacier flux on annual timescales because subglacial drainage networks have evolved to drain away the additional water inputs (e.g., Sole et al., 2013; Tedstone et al., 2015; Stevens et al., 2016; Nienow et al., 2017; King et al., 2018).

3.3.1.5 Drivers of ice sheet mass change

3.3.1.5.1 Ocean drivers

The reduction of ice-shelf buttressing that has dominated AIS mass loss (Section 3.3.1.2) has been driven primarily by increases in sub-ice-shelf melting (Khazendar et al., 2013; Pollard et al., 2015; Cook et al., 2016; Rintoul et al., 2016; Walker and Gardner, 2017; Adusumilli et al., 2018; Dow et al., 2018a; Minchew et al., 2018) (*high confidence*). Shoaling of relatively warm Circumpolar Deep Water has controlled recent variability in melting in the Amundsen and Bellingshausen seas, Wilkes Land (Roberts et al., 2018) and the AP (*medium confidence*) (Jacobs et al., 2011; Pritchard et al., 2012; Depoorter et al., 2013; Rignot et al., 2013; Dutrieux et al., 2014; Paolo et al., 2015; Wouters et al., 2015; Christianson et al., 2016; Cook et al., 2016; Jenkins et al., 2018; Roberts et al., 2018). Changes in winds have driven this shoaling by affecting continental-shelf-edge undercurrents (Walker et al., 2013; Dutrieux et al., 2014; Kimura et al., 2017) and overturning in coastal polynyas (St-Laurent et al., 2015; Webber et al., 2017) (*medium confidence*). Winds over the Amundsen Sea are highly variable, however, with complex interactions between the Southern Annular Mode (SAM), El Niño/Southern Oscillation (ENSO), Atlantic Multidecadal Oscillation, and the Amundsen Sea Low (Uotila et al., 2013; Li et al., 2014; Turner et al., 2016) (SM3.1.3).

Through their effects on Antarctic coastal ocean circulation, ENSO or other tropical-ocean variability may have triggered changes to Pine Island Glacier in the 1940s (Smith et al., 2017c) and again in the 1970s and 1990s (Jenkins et al., 2018), and recent ENSO variability is correlated with recent changes in ice-shelf thickness (Paolo et al., 2018) (*medium confidence*). Such coupling between wind variability, ocean upwelling, ice-shelf melt, buttressing and glacier flow rate has also been observed in EAIS, at Totten Glacier, Wilkes Land (Greene et al., 2017).

Around Greenland, 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) (*medium confidence*). Water temperatures near the grounding zone of GIS outlet glaciers are critically important to their calving rate (O'Leary and Christoffersen, 2013) (*medium confidence*), and warm waters have been observed interacting with major GIS outlet glaciers (*high confidence*) (e.g., Holland et al., 2008; Straneo et al., 2017).

The processes behind warm-water incursions in coastal Greenland that force glacier retreat remain unclear, however (Straneo et al., 2013; Xu et al., 2013b; Bendtsen et al., 2015; Murray et al., 2015; Cowton et al., 2016; Miles et al., 2016), and there is *low confidence* in understanding coastal GIS glacier response to ocean forcing because submarine melt rates, calving rates (Rignot et al., 2010; Todd and Christoffersen, 2014; Benn et al., 2017), bed and fjord geometry, and the roles of ice melange and subglacial discharge (Enderlin et al., 2013; Gladish et al., 2015; Slater et al., 2015; Morlighem et al., 2016; Rathmann et al., 2017) are poorly understood, and extrapolation from a small sample of glaciers is impractical (Moon et al., 2012; Carr et al., 2013; Straneo et al., 2016; Cowton et al., 2018).

3.3.1.5.2 Atmospheric drivers

Snow accumulation and surface melt in Antarctica are influenced by the Southern Hemisphere extratropical circulation (SM3.1.3), which has intensified and shifted poleward in austral summer from 1950-2012 (Arblaster et al., 2014; Swart et al., 2015a) (*medium confidence*). 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), and from 1979-2013 has contributed to intensified atmospheric circulation, increasing and decreasing snowfall in the western and eastern AP respectively (Marshall et al., 2017) (*medium confidence*). WAIS accumulation trends (Section 3.3.1.2) resulted from a deepening of the Amundsen Sea Low over recent decades (Raphael et al., 2016) (*high confidence*).

During the 1990s, WAIS experienced record surface warmth relative to the past 200 years, though similar conditions occurred for 1% of the preceding 2000 years (Steig et al., 2013), and WAIS surface melting remains limited. In contrast, AP surface melting has intensified since the mid-20th century and the last three decades were unprecedented over 1000 years (Abram et al., 2013a). The northeast AP began warming 600 years ago and past-century rates were unusual over 2000 years (Mulvaney et al., 2012b; Stenni et al., 2017). Increased föhn winds due to the more positive SAM (Cape et al., 2015) caused increased surface melting on the Larsen ice shelves (Grosvenor et al., 2014; Luckman et al., 2014; Elvidge et al., 2015) and after 11,000 years intact, the 2002 melt-driven collapse of the Larsen B ice shelf followed strong warming between the mid–1950s and the late 1990s (Domack et al., 2005) (*medium confidence*).

In Greenland, associations between atmospheric pressure indices such as the North Atlantic Oscillation (NAO) and temperature, insolation and snowfall indicate with *high confidence* that, as in Antarctica, variability of large-scale atmospheric circulation is an important driver of SMB changes (Fettweis et al., 2013; Tedesco et al., 2013; Ding et al., 2014; Tedesco et al., 2016b; Ding et al., 2017; Hofer et al., 2017). A post-1990s decrease in summer NAO reflects increased anticyclonic weather (e.g., Tedesco et al., 2013; Hanna et al., 2015) that advected warm air over the GIS, explaining ~70% of summer surface warming from 2003-2013 (Fettweis et al., 2013; Tedesco et al., 2013; Mioduszewski et al., 2016), and reduced the cloud cover, increasing shortwave insolation (Tedesco et al., 2013) that, combined with albedo feedbacks (Box et al., 2012; Charalampidis et al., 2015; Tedesco et al., 2016a; Stibal et al., 2017; Ryan et al., 2018) (*high confidence*), explains most of the post-1990s melt increase (Hofer et al., 2017). These drivers culminated in July 2012 in exceptional warmth and surface melt up to the ice sheet summit (Nghiem et al., 2012; Tedesco et al., 2013; Hanna et al., 2014; Hanna et al., 2016; McLeod and Mote, 2016).

3.3.1.6 Natural and Anthropogenic Forcing

There is *medium agreement* but *limited evidence* of anthropogenic forcing of AIS mass balance through both SMB and glacier dynamics (*low confidence*). Partitioning between natural and human drivers of atmospheric and ocean circulation changes remains very uncertain. Partitioning is challenging because, along with the effects of greenhouse gas increases and stratospheric ozone depletion (Waugh et al., 2015; England et al., 2016; Li et al., 2016a), atmospheric and ocean variability in the areas of greatest AIS mass change are affected by a complex chain of processes (e.g., Fyke et al., 2018; Zhang et al., 2018a) that exhibit considerable natural variability and have multiple interacting links to sea surface conditions in the Pacific (Schneider et al., 2015; England et al., 2016; Raphael et al., 2016; Clem et al., 2017; Steig et al., 2017; Paolo et al., 2018) and Atlantic (Li et al., 2014), with additional local feedbacks (e.g., Stammerjohn et al., 2012; Goosse and Zunz, 2014). Recent AP warming and consequent ice-shelf collapses have evidence of a link to anthropogenic ozone and greenhouse-gas forcing via the SAM (e.g., Marshall, 2004; Shindell, 2004; Arblaster and Meehl, 2006; Marshall et al., 2006; Abram et al., 2014) and to anthropogenic Atlantic sea-surface warming via the Atlantic Multidecadal Oscillation (e.g., Li et al., 2014). This warming was highly unusual over the last 1000 years but not unprecedented, and along with subsequent cooling is within the bounds of the large natural decadal-scale climate variability in this region (Mulvaney et al., 2012a; Turner et al., 2016). More broadly over the AP and coastal WAIS where dynamic mass losses are concentrated, natural variability in atmospheric and ocean forcing appear to dominate observed mass balance (*medium confidence*) (Smith and Polvani, 2017; Jenkins et al., 2018).

Evidence exists for an anthropogenic role in the atmospheric circulation (NAO) changes that have driven GIS mass loss (Section 3.3.1.5.2) (*medium confidence*), although this awaits formal attribution testing (e.g., Easterling et al., 2016). Arctic amplification of anthropogenic warming (e.g., Serreze et al., 2009) affects atmospheric circulation (Francis and Vavrus, 2015; Mann et al., 2017) and has reduced sea-ice extent (Section 3.2.1.1.1), feeding back to exacerbate both warming and NAO changes (Screen and Simmonds, 2010) that impact GIS mass balance. Negative-NAO wind patterns increased GIS melt observed in a 40-year runoff signal (Ahlstrom et al., 2017), and an increase in melting beginning in the mid-1800s closely followed the onset of industrial-era Arctic warming and emerged beyond the range of natural variability in the last few decades (Graeter et al., 2018; Trusel et al., 2018) (Section 3.3.1.4).

3.3.1.7 Ice sheet projections

Section 4.2 assesses the sea level impacts from observed and projected changes in ice sheets.

3.3.2 Polar Glacier Changes

3.3.2.1 Observations, Components of Change, and Drivers

Chapter 3 assesses changes in polar glaciers in the Canadian and Russian Arctic, Svalbard, Greenland and Antarctica, independent of the Greenland and Antarctic ice sheets (Figure 3.8). Glaciers in all other regions including Alaska, Scandinavia and Iceland are assessed in Chapter 2.

Changes in the mass of Arctic glaciers for the 'present day' (2006 to 2015) are assessed using a combination of satellite observations and direct measurements (Figure 3.8; Appendix 2.A, Table 1). During this period, glacier mass loss was largest in the periphery of Greenland (-47 ± 16 Gt yr^{-1}), followed by Arctic Canada North (-39 ± 8 Gt yr^{-1}), Arctic Canada South (-33 ± 9 Gt yr^{-1}), the Russian Arctic (-15 ± 12 Gt yr^{-1}) and Svalbard and Jan Mayen (-9 ± 5 Gt yr^{-1}). When combined with the Arctic regions covered in Chapter 2 (Alaska, the Yukon territory of Canada, Iceland and Scandinavia), Arctic glaciers as a whole lost mass at a rate of -213 ± 29 Gt yr^{-1}, a sea level contribution of 0.59 ± 0.08 mm yr^{-1} (*high confidence*). Overall during this period, Arctic glaciers caused a similar amount of sea level rise to the Greenland Ice Sheet (Section 3.3.1.3), but their rate of mass loss per unit area was larger (Bolch et al., 2013).

There is *limited evidence* (*high agreement)* that the current rate of glacier mass loss is larger than at any time during the past 4000 years (Fisher et al., 2012; Zdanowicz et al., 2012). Further back in time during the early- to mid- Holocene, pre-historic glacial deposits, ice core records, and numerical modelling evidence shows that many Arctic glaciers were at various stages similar to or smaller than present (Gilbert et al., 2017; Zekollari et al., 2017), experienced greater melt rates (Lecavalier et al., 2017), or may have disappeared altogether (Solomina et al., 2015) (*medium confidence*). This evidence, however, does not provide a complete assessment of the rates and magnitudes of past glacier mass loss.

Atmospheric circulation changes (Box et al., 2018) have led to pan-Arctic variability in glacier mass balance (*high confidence*), including different rates of retreat between eastern and western glaciers in Greenland's periphery (Bjørk et al., 2018), and a high rate of surface melt in the Canadian Arctic (Gardner et al., 2013; Van Wychen et al., 2016; Millan et al., 2017) through persistently high summer air temperatures (Bezeau et al., 2014; McLeod and Mote, 2016). Atmospheric circulation anomalies from 2007-2012 associated with glacier mass loss are also linked to enhanced Greenland ice sheet melt (Section 3.3.1.4) and Arctic sea ice loss (Section 3.2.1.1), and exceed by a factor of two the interannual variability in daily mean pressure (sea level and 500 hPa) of the Arctic region over the 1871-2014 period (Belleflamme et al., 2015) (Section 3.3.1.6).

Increased surface melt on Arctic glaciers has led to a positive feedback from lowered surface albedo, causing further melt (Box et al., 2012), and in Svalbard, mean glacier albedo has reduced between 1979 and 2015 (Möller and Möller, 2017). Across the Arctic, increased surface melt and subsequent ice-layer formation via refreezing within snow and firn also reduces the ability of glaciers to store meltwater, increasing runoff (Zdanowicz et al., 2012; Gascon et al., 2013a; Gascon et al., 2013b; Noël et al., 2017; Noël et al., 2018).

Between the 1990s and 2017, tidewater glaciers have exhibited regional patterns in glacier dynamics; glaciers in Arctic Canada have largely decelerated, while glaciers in Svalbard and the Russian Arctic have accelerated (Van Wychen et al., 2016; Strozzi et al., 2017). 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). Acceleration due to surging (an internal dynamic instability) of a few key glaciers has dominated dynamic ice discharge on time-scales of years to decades (Van Wychen et al., 2014; Dunse et al., 2015).

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 terminus thinning and response to warmer ocean temperatures (McMillan et al., 2014a) (*low confidence*). Iceberg calving rates in Svalbard are linked to ocean temperatures which control rates of submarine melt (Luckman et al., 2015; Vallot et al., 2018) (*medium confidence*). Rapid disintegration of ice shelves in the Canadian and Russian Arctic continues and has led to acceleration and thinning in tributary-glacier basins (*high confidence*) (Willis et al., 2015; Copland and Mueller, 2017).

Little information is available on Holocene and historic changes in glaciers in Antarctica (separate from the ice sheet), and on sub-Antarctic islands (Hodgson et al., 2014). Mass changes of glaciers in these regions between 2006 and 2015 ($–90±860$ Gt yr$^{–1}$) have *low confidence* as they are based on a single data compilation with large uncertainties in the Antarctic region (Zemp et al., 2019) (Figure 3.8). *Limited evidence* with *high agreement* from individual glaciers suggests that regional variability in glacier mass changes may be linked to changes in the large-scale Southern Hemisphere atmospheric circulation (Section 3.3.1.5.2). On islands adjacent to the Antarctic Peninsula, glaciers experienced retreat and mass loss during the mid to late 20th Century, but since around 2009 there has been a reduction in mass loss rate or a return to slightly positive balance (Navarro et al., 2017; Oliva et al., 2017). Reduced mass loss has been linked to increased winter snow accumulation and decreased summer melt at these locations, associated with recent deepening of the circumpolar pressure trough (Oliva et al., 2017). Conversely, on the sub-Antarctic Kerguelen Islands, increased glacier mass loss (Verfaillie et al., 2015) may be due to reduced snow accumulation rather than increased air temperature as a result of southward migration of storm tracks (Favier et al., 2016).

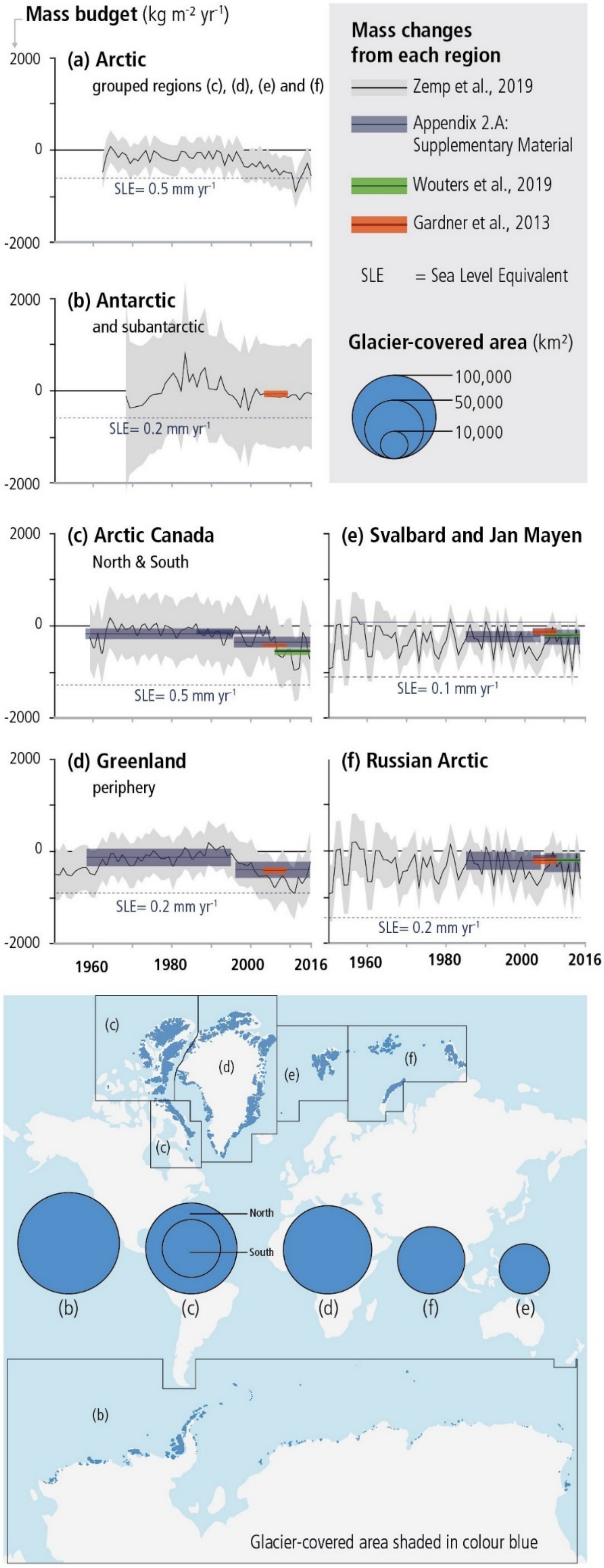


Figure 3.8: Glacier mass budgets for the six polar regions assessed in Chapter 3. Glacier mass budgets for all other regions (including Iceland, Scandinavia and Alaska) are shown in Chapter 2, Figure 2.4. Regional time series of annual mass change are based on glaciological and geodetic balances (Zemp et al., 2019). Superimposed are multi-year

averages by Wouters et al. (2019) and Gardner et al. (2013) from the Gravity Recovery and Climate Experiment (GRACE). Estimates by Gardner et al. (2013) were used in AR5. Additional regional estimates in some regions are listed in Appendix 2.1, Table 1. Annual and time-averaged mass-budget estimates include the errors reported in each study. Glacier outlines and areas are based on RGI Consortium (2017).

3.3.2.2 Projections

Projections of all glaciers, including those in polar regions, are covered in Cross-Chapter Box 6 in Chapter 2.

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 8 in Chapter 3).

3.3.3.2 Physical Oceanography

The major large-scale impacts of freshwater release 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 (Section 6.7). Freshwater release also affects local circulation within fjords through two principle mechanisms; subglacial release from tidewater glaciers enhances buoyancy driven circulation, whereas runoff from land-terminating glaciers contributes to surface layer freshening and estuarine circulation (Straneo and Cenedese, 2015). There is *limited evidence* that freshening occurred between 2003-2015 in North East Greenland fjords and coastal waters (Sejr et al., 2017).

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* in 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 Gt y^{-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 the sparsity of information 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 *low confidence* in how stratification changes are affecting sea ice and basal ice shelf melt. An increase in stratification caused by release 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; Bronselaer et al., 2018; Purich et al., 2018), 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) (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

Both polar ice sheets have the potential to release dissolved and sediment-bound nutrients and organic carbon directly to the surface ocean via subglacial and surface meltwater, icebergs, melting of the base of ice shelves (Shadwick et al., 2013; Wadham et al., 2013; Hood et al., 2015; Herraiz-Borreguero et al., 2016; Raiswell et al., 2016; Yager et al., 2016; Hodson et al., 2017), in addition to indirectly stimulating nutrient input via upwelling associated with subglacial meltwater plumes (Meire et al., 2016b; Cape et al., 2018; Hopwood et al., 2018; Kanna et al., 2018) (Figure 3.9). These nutrient additions stimulate primary production in the surrounding ocean waters in some regions (*medium confidence*) (Gerringa et al., 2012; Death et al., 2014; Duprat et al., 2016; Arrigo et al., 2017b). There is also some evidence to support melting ice sheets as source of contaminants (AMAP, 2015).

In Greenland, direct measurements suggest that meltwater is a significant source of bioavailable silica and iron (Bhatia et al., 2013; Hawkings et al., 2014; Meire et al., 2016a; Hawkings et al., 2017) but may be less important for the supply of bioavailable forms of dissolved nitrogen or phosphorous (Hawkings et al., 2016; Wadham et al., 2016), which often limit the integrated primary production during summer in fjords (Meire et al., 2016a; Hopwood et al., 2018). The offshore export of iron, however, has been linked to primary productivity in surface ocean waters in the Labrador Sea (Arrigo et al., 2017b) (*limited evidence, high agreement*).

Subglacial meltwater plumes from tidewater glaciers have emerged recently as an important indirect source of nutrients to fjords, by entraining nutrient-replete seawater (Meire et al., 2016b; Meire et al., 2017; Cape et al., 2018; Hopwood et al., 2018; Kanna et al., 2018) (*medium evidence, high agreement*). There is *medium evidence* with *high agreement* that these upwelled nutrient fluxes enhance primary production in fjords over a distance of up to 100 km along the trajectory of the outflowing plume (Juul-Pedersen et al., 2015; Cape et al., 2018; Kanna et al., 2018).

In Antarctica**,** there is *medium evidence* with *high agreement* that enhanced input of iron from ice shelves, glacial meltwater and icebergs stimulates primary production in polynyas, coastal regions and the wider Southern Ocean (Gerringa et al., 2012; Shadwick et al., 2013; Herraiz-Borreguero et al., 2016). Satellite observations and modelling also indicate variable potential for icebergs to fertilise the Southern Ocean beyond the coastal zone (Death et al., 2014; Duprat et al., 2016; Wu and Hou, 2017).

Dissolved nutrient fluxes from ice sheets may be increasing during high melt years (Hawkings et al., 2015). The dominant sediment-bound fraction, however, may not increase with rising melt (Hawkings et al., 2015). Thus, there is *low confidence* overall in the magnitude of the response of direct nutrient fluxes from ice sheets to enhanced melting.

Future predictions of nutrient cycling proximal to ice sheets is made more challenging by the landward progression of marine-terminating glaciers and the 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). The erosion of newly-exposed glacial sediments in front of retreating land-terminating glaciers (Monien et al., 2017) and changes in the diffuse nutrient fluxes from newly exposed glacial sediments on the seafloor (Wehrmann et al., 2014) may amplify nutrient supply, whilst other nutrient sources may be cut off (e.g., icebergs, upwelling of marine water; Meire et al., 2017) (*low confidence*).

There is *medium evidence* with *high agreement* 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 upwelling a source of nutrients, thereby reducing summer productivity in Greenland fjord ecosystems (Meire et al., 2017; Hopwood et al., 2018).

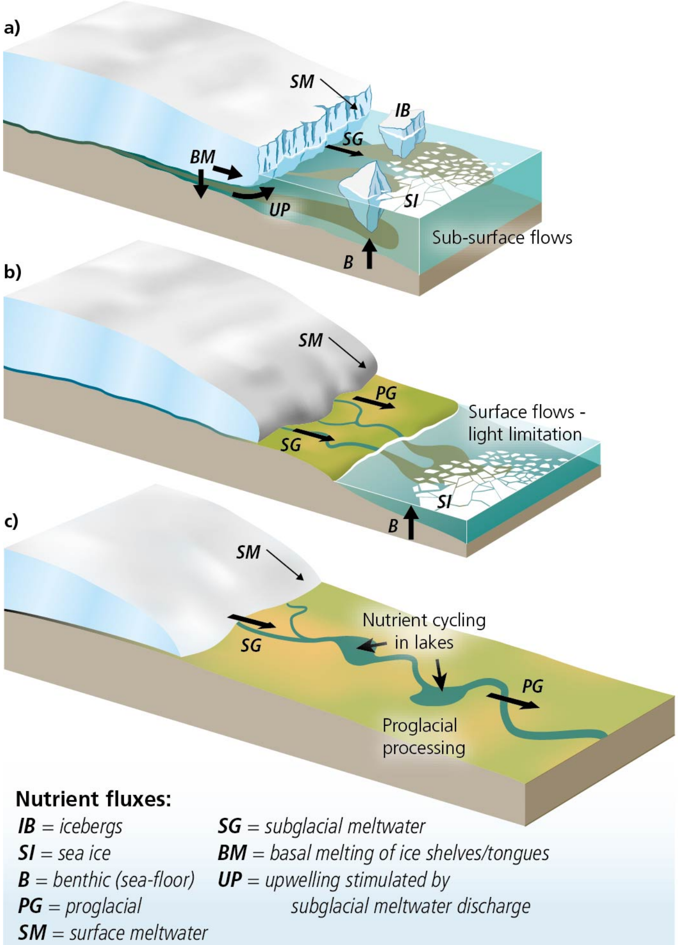


Figure 3.9. Potential shifts in nutrient fluxes with landward retreat of marine-terminating glaciers (a) at different stages (b and c).

3.3.3.4 Ecosystems

For Greenland and Svalbard, there is *limited evidence* with *high agreement* 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). 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). There is also evidence that marine-terminating glaciers are important feeding areas for marine mammals and seabirds at Greenland (Laidre et al., 2016) and Svalbard (Lydersen et al., 2014).

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; Ingels et al., 2018). 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) (Section 3.2.3.2.1), and have resulted in enhanced carbon uptake by coastal marine ecosystems (*medium confidence*), although quantitative estimates of biological carbon uptake are highly variable (Trathan et al., 2013; Barnes et al., 2018). Newly-available habitat on coastlines may also provide breeding or haulout 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 abundance and biodiversity of benthic macro-organisms (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 8 HERE]

Cross-Chapter Box 8: Future Sea Level Changes and Marine Ice Sheet Instability

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

Over the last century, glaciers were the main contributors to increasing ocean water mass (Section 4.2.1.2). However, most terrestrial frozen water is stored in Antarctic and Greenland ice sheets, and future changes in their dynamics and mass balance will cause sea level rise over the 21st century and beyond (Section 4.2.3).

About a third of Antarctic Ice Sheet (AIS) is 'marine ice sheet', i.e. rests on bedrock 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 - so-called Marine Ice Sheet and Marine Ice Cliff instabilities, as discussed below.

In many places around the AIS margin, the seaward-flowing ice forms floating ice shelves (Figure CB8.1). 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 (Section 3.3.1.2).

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 (i.e., flowing on the glacier surface) 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.

The future dynamic 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 CB8.1a). This self-sustaining process is known as Marine Ice Sheet Instability (MISI). 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).

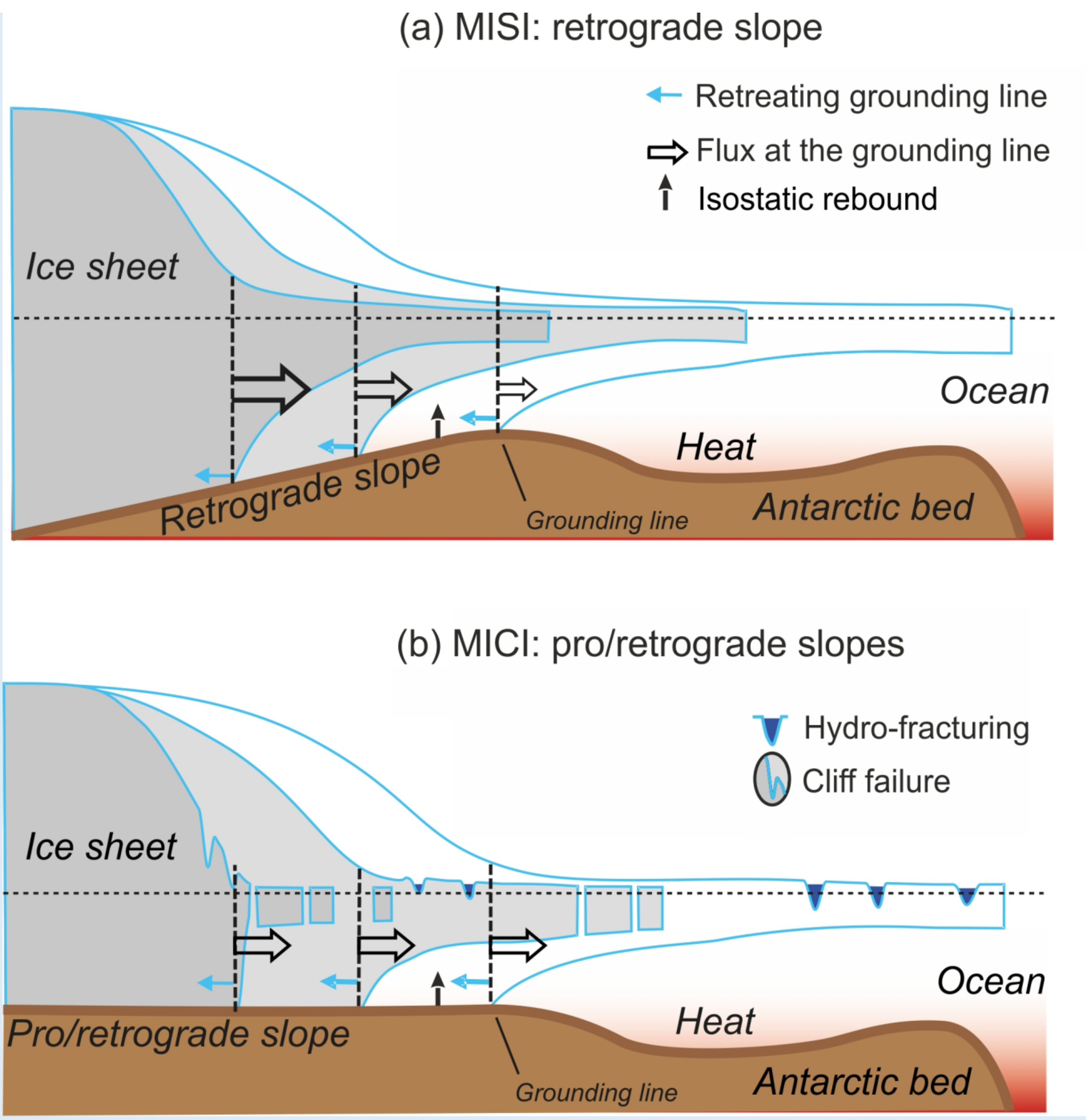


Figure CB8.1: Schematic representation of Marine Ice Sheet Instability (MISI, a) and Marine Ice Cliff Instability (MICI, b) from Pattyn (2018). (a) thinning of the buttressing ice shelf leads to acceleration of the ice sheet flow and thinning of the marine-terminated ice margin. Because bedrock under 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 (at least ~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 been shown recently (Barletta et al., 2018) that the Amundsen Sea Embayment experiences unexpectedly fast bedrock uplift (up to 41 mm yr^{-1}, due to mantle viscosity much lower than the global average) 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., 2015b). Totten's current behaviour suggests that East Antarctica could become a substantial contributor to future sea level rise, as it has been in the previous warm periods (Aitken et al., 2016). It is not clear, however, if the changes observed recently are a linear response to increased ocean forcing (Section 3.3.1.2), or an indication that MISI has commenced (Roberts et al., 2018).

The disappearance of ice shelves may allow the formation of ice cliffs, which may be inherently unstable if they are tall enough (subaerial cliff height between 100 and 285 m) to generate stresses that exceed the strength of the ice (Parizek et al., 2019). This ice cliff failure can lead to ice sheet retreat via a process called marine ice cliff instability (MICI; Figure CB8.1b), 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 (Section 4.2.3.1.2). Overall, there is *low agreement* on the exact MICI mechanism and *limited evidence* of its occurrence in the present or the past. Thus the potential of MICI to impact the future sea level remains very uncertain (Edwards et al., 2019).

Limited evidence from geological records and ice sheet modelling suggests that parts of AIS experienced rapid (i.e., on centennial time-scale) retreat *likely* due to ice sheet instability processes between 20,000 and 9,000 years ago (Golledge et al., 2014; Weber et al., 2014; Small et al., 2019). Both the West (including Pine Island glacier) and the East Antarctic Ice Sheet 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 (Section 3.3.1.5.1) (Golledge et al., 2014; Hillenbrand et al., 2017) and MISI (Jones et al., 2015b). *Limited 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 (Section 4.2.3.2) due to poor understanding of mechanisms and lack of observational data on bed topography, isostatic rebound rates, etc. 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) (Section 4.2.3.1.2).

The Greenland Ice Sheet has limited direct access to the ocean through relatively narrow subglacial troughs (Morlighem et al., 2017), and most of the bedrock at the ice-sheet margin is above sea level (Figure 4.5). However, since AR5 it has been argued 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 that Greenland was nearly ice free for extensive episodic periods during the Pleistocene, suggesting a sensitivity to deglaciation under climates similar to or slightly warmer than present (Schaefer et al., 2016).

A MICI-style behaviour is seen today in Greenland at the termini of Jakobshavn and Helheim glaciers (Parizek et al., 2019), 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, this assessment finds that unstable retreat and thinning of some Antarctic glaciers, and to a lesser extent Greenland outlet glaciers, may be underway. However, the timescale and future rate of these processes is not well known, casting deep uncertainty on projections of the sea level contributions from the Antarctic ice sheet (Cross-Chapter Box 5 in Chapter 1, Section 4.2.3.1).

[END CROSS-CHAPTER BOX 8 HERE]

3.4 Arctic Snow, Freshwater Ice and Permafrost: 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 are almost 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 (land areas north of 60°N) spring snow cover extent have occurred since satellite charting began in 1967 (Estilow et al., 2015). Declines in May and June of –3.5% (± 1.9%) and –13.4% respectively per decade (± 5.4%) between 1981 and 2018 (relative to the 1981–2010 mean) were determined from multiple datasets based on the methodology of (Mudryk et al., 2017) (Figure 3.10) (*high confidence*). The loss of spring snow extent is reflected in shorter snow cover duration estimated from surface observations (Bulygina et al., 2011; Brown et al., 2017), satellite data (Wang et al., 2013; Estilow et al., 2015; Anttila et al., 2018), 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., 2017). 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., 2017) (*high confidence*). There is *low confidence* in positive October and November snow cover extent trends apparent in a single dataset (Hernández-Henríquez et al., 2015) because 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 between 1966 and 2014 (Bulygina et al., 2011; Osokin and Sosnovsky, 2014). There is *medium confidence* in this trend because the pointwise nature of these measurements does 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., 2017). The timing of maximum snow depth has shifted earlier by 2.7 days per decade for the North American Arctic (Brown et al., 2017); 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., 2017). 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., 2017) 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 (*high confidence*). 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., 2017), and 700,000 to 800,000 km^2 lost in autumn (Derksen and Brown, 2012; Brown and Derksen, 2013) (*high confidence*).

There is *high confidence* that darkening of snow through the deposition of black carbon and other light absorbing particles enhances snow melt (Bullard et al., 2016; Skiles et al., 2018; Boy et al., 2019). The global direct radiative forcing for black carbon in seasonal snow and over 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 black carbon 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 black carbon emissions and reductions in dust darkening (*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 black carbon 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*).

Precipitation remains a sparse and highly uncertain measurement over Arctic land areas: *in situ* datasets remain uncertain (Yang, 2014) and are largely regional (Kononova, 2012; Vincent et al., 2015). Atmospheric reanalyses show increases in Arctic precipitation in recent decades (Lique et al., 2016; Vihma et al., 2016), but 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). Despite improved process understanding, estimates of sublimation loss during blowing snow events remain a key uncertainty in the mass budget of the Arctic snowpack (Sturm and Stuefer, 2013).

3.4.1.2 Permafrost

3.4.1.2.1 Temperature

Record high temperatures at ~10–20 m depth in the permafrost (near or below the depths affected by intra-annual fluctuation in temperature) have been documented at many long-term monitoring sites in the Northern Hemisphere circumpolar permafrost region (AMAP, 2017d) (Figure 3.10) (*very high confidence*). At some locations, the temperature is 2–3°C higher than 30 years ago. During the decade between 2007 and 2016, the rate of increase in permafrost temperatures was 0.39 ± 0.15°C for colder continuous zone permafrost monitoring sites and 0.20 ± 0.10°C for warmer discontinuous zone permafrost (Biskaborn et al., 2019). Relatively smaller increases in permafrost temperature in warmer sites indicate that permafrost is thawing with heat absorbed by the ice-to-water phase change, and as a result, the active layer may be increasing in thickness. In contrast to temperature, there is only *medium confidence* that active layer thickness across the region has increased. This confidence level 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 permafrost in some cases because the surface subsides when ground ice melts and drains (Romanovsky et al., 2016; AMAP, 2017d; Streletskiy et al., 2017). Permafrost in the Southern Hemisphere polar region occurs in ice-free exposed areas (Bockheim et al., 2013), 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 terrestrial permafrost region (Gruber, 2012). Antarctic permafrost temperatures are generally colder (Noetzli et al., 2017) and increased 0.37 ± 0.10°C between 2007 and 2016 (Biskaborn et al., 2019).

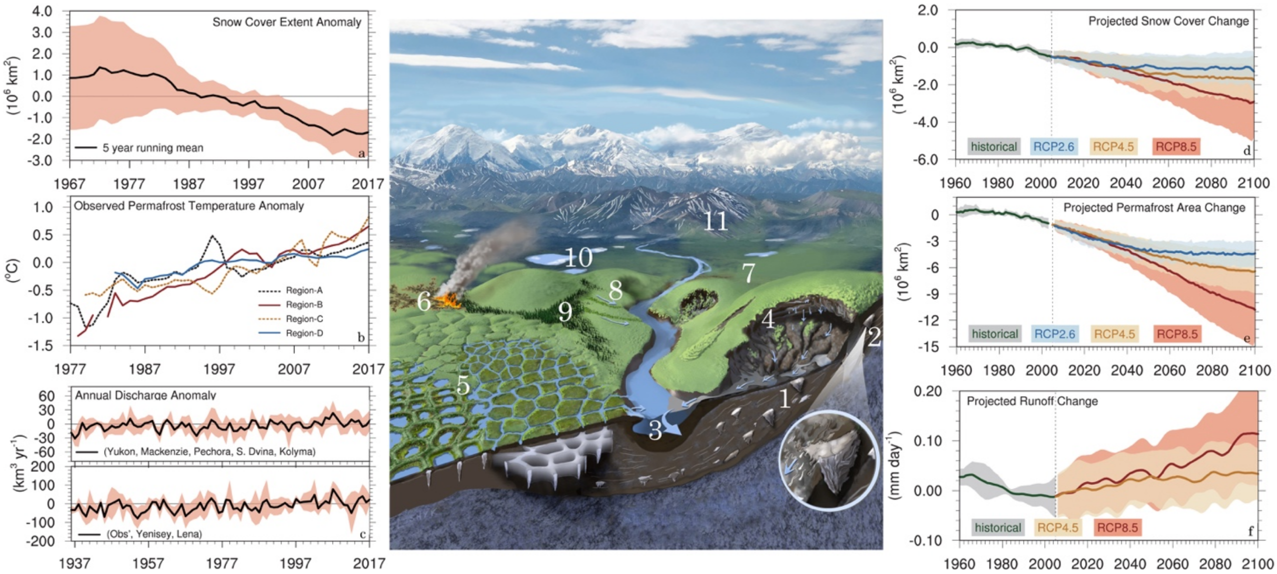


Figure 3.10: Schematic of important land surface components influenced by the Arctic terrestrial cryosphere (centre): permafrost (1); ground ice (2); river discharge (3); abrupt thaw (4); surface water (5); fire (6); tundra (7); shrubs (8); boreal forest (9); lake ice (10); seasonal snow (11). Left column: time series of snow cover extent anomalies in June (relative 1981–2010 climatology) from 5 products based on the approach of Mudryk et al. (2017); permafrost temperature change normalized to a baseline period (Romanovsky et al., 2017) and runoff from northern flowing watersheds normalized to a baseline period (1981–2010) (Holmes et al., 2018). Right column: CMIP5 multi-model average for different Representative Concentration Pathway scenarios for June snow cover extent (based on Thackeray et al., 2016), area of near-surface permafrost, and runoff to the Arctic Ocean (based on McGuire et al., 2018).

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 (Kanevskiy et al., 2013; Raynolds et al., 2014). Excess ice in permafrost is typical, varying 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 areas where impacts of thaw could be greatest include the Yedoma deposits in Siberia, Alaska, and the Yukon in Canada, with ice divided between massive wedges interspersed with frozen soil/sediment containing pore ice and smaller ice features (Schirrmeister et al., 2011; Strauss et al., 2017). Other areas including for example Northwestern Canada, the Canadian Archipelago, 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; Coulombe et al., 2019). The location and volume of ground ice integrated across the northern permafrost region (5.63-36.55x10^3 km^3, equivalent to 2-10 cm sea level rise) is known with *medium confidence* and with no recent updates at the circumpolar scale (Zhang et al., 2008).

3.4.1.2.3 Carbon

The permafrost region 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 CO_2 and methane as permafrost thaws in a warming climate, thus accelerating the pace of climate change (Schuur et al., 2015). The current best mean estimate of total (surface plus deep) organic soil carbon (terrestrial) in the northern circumpolar permafrost region (17.8x10^6 km^2 area) is 1460 to 1600 petagrams (*medium confidence*) (Pg; 1 Pg = 1 billion metric tons) (Schuur et al., 2018a). All permafrost-region soils estimated to 3 m in depth (surface) contain 1035 ± 150 Pg C (Tarnocai et al., 2009; Hugelius et al., 2014) (*high confidence*). Of the carbon in the surface, 800–1000 Pg C is perennially frozen, with the remainder contained in seasonally-thawed soils. The northern circumpolar permafrost region occupies only 15% of the total global soil area, but the 1035 Pg C adds another 50% to the rest of the 3 m soil carbon inventory (2050 Pg C for all global biomes excluding tundra and boreal; Jobbágy and Jackson, 2000; 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) and 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 degrading slowly 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 below sea level depth) that was formerly exposed as land (Ruppel et al., 2016). These observations are supported by similar studies in the Siberian Arctic Seas (Portnov et al., 2013), and by modelling that suggests that subsea permafrost would be thawed many meters below the seabed under current submerged conditions (Anisimov et al., 2012; AMAP, 2017d; Angelopoulos et al., 2019).

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 discussed in Section 3.4.1.2.1. Climate change can also 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 (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 (Kanevskiy et al., 2013).

Pulse disturbances often rapidly remove the insulating soil organic layer, leading to permafrost degradation (Gibson et al., 2018). 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; Hanes et al., 2019). 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 (Gillett, 2004; Veraverbeke et al., 2017). 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. The annual area burned in Arctic tundra is generally small compared to the forested boreal biome. In Alaska – the only region where estimates of burned area exist for both boreal forest and tundra vegetation types – tundra burning averaged approximately 270 km^2 per year during the last half century (French et al., 2015), accounting for 7% of the average annual area burned throughout the state (Pastick et al., 2017). There is *high confidence* that changes in the fire regime are degrading permafrost faster than had occurred over the historic successional cycle (Turetsky et al., 2011; Rupp et al., 2016; Pastick et al., 2017), and that the effect of this driver of permafrost change is under-represented in the permafrost temperature observation network.

Abrupt permafrost thaw occurs when changing environmental and ecological conditions interact with geomorphological processes. Melting ground ice causes the ground surface to subside. Pooling or flowing water causes localized permafrost thaw and sometimes mass erosion. 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 process 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 *medium 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 (Kokelj et al., 2017; Nitze et al., 2018). New research at the global scale has revealed that $3.6x10^6$ km^2, about 20% of the northern permafrost region, appears to be vulnerable to abrupt thaw (Olefeldt et al., 2016).

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 historical 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 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 Runoff and surface water

A general trend of increasing discharge has been observed for large Siberian (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 (*medium confidence*). Between 1976 and 2015, trends are 3.3 ± 1.6% for Eurasian rivers and 2.0 ± 1.8% for North American rivers (Holmes et al., 2018) (Figure 3.10). Extreme regional runoff events have also been identified (Stuefer et al., 2017). An observed increase in baseflow in the North American (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009) and Eurasian Arctic (Smith et al., 2007; Duan et al., 2017) over the last several decades is attributable to permafrost thaw and concomitant enhancement in groundwater discharge. 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 over Siberian regions (Liu et al., 2005; Lammers et al., 2007). Increases in discharge and water temperature in the spring season represent notable freshwater and heat fluxes to the Arctic Ocean (Yang et al., 2014).

A large proportion of low-lying Arctic land areas are covered by lakes 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; Ulrich et al., 2017; Pastick et al., 2018). In landscapes with degrading ice-wedge polygons, 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). Enhanced subsurface connectivity from thaw in discontinuous permafrost serves tempers short-term lake fluctuations (Rey et al., 2019).

3.4.1.3.3 Drivers

There is *high confidence* that environmental drivers of Arctic surface water change are diverse and depend on local and regional factors such as permafrost properties and geomorphology (Nitze et al., 2018). Thermokarst lake expansion has been observed in the continuous permafrost of northern Siberia (Smith et al., 2005; Polishchuk et al., 2015) and Alaska (Jones et al., 2011); surface water area reduction has been observed in discontinuous permafrost of central and southern Siberia (Smith et al., 2005; Sharonov, 2012), western Canada (Labrecque et al., 2009; Carroll et al., 2011; Lantz and Turner, 2015) and interior Alaska (Chen et al., 2012; Rover et al., 2012). 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.

Satellite and model-derived estimates of evapotranspiration show increases across the Arctic (Rawlins et al., 2010; Liu et al., 2014; Liu et al., 2015b; Fujiwara et al., 2016; Suzuki et al., 2018) (*medium confidence*). 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).

3.4.2 Projections

3.4.2.1 Seasonal Snow

Historical simulations from CMIP5 models tend to underestimate observed reductions in 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., 2016b), 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., 2017) driven by increased surface temperature over essentially all Arctic land areas (Hartmann et al., 2013). There is *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. Under RCP4.5, Arctic snow cover duration stabilizes at 5–10% reduction (compared to a 1986–2005 reference period); under RCP8.5, snow cover duration declines reach –15 to –25% (Brown et al., 2017) (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., 2017) (*medium confidence*). Projected snow water equivalent increases across the North American Arctic are only modest, emerge later in the century, and only under RCP8.5 (Brown et al., 2017). These projected increases are due to enhanced snowfall (Krasting et al., 2013) from a more moisture-rich Arctic atmosphere coupled with winter season temperatures that remain sufficiently low for precipitation to fall as snow. There is *low confidence* in changes to snow properties such as density and stratigraphy (relevant for understanding the impacts of changes to Arctic snow on ecosystems) which are not resolved directly by climate model simulations, but require detailed snow physics models.

3.4.2.2 Permafrost

Circumpolar- or global-scale models represent permafrost degradation in response to warming scenarios as increases in thaw depth only. The CMIP5 models project with *high confidence* that thaw depth 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: 2-66% for RCP2.6, 15–87% under RCP4.5 and 30–99% under RCP8.5. A more recent analysis of near-surface permafrost trends from a subset of models that self-identified as structurally representing the permafrost region had a significantly smaller range of estimated present day near-surface permafrost area (13.1–19.3x10^6 km^2; mean ± SD, 14.1 ± 3.5 x 10^6 km^2) (McGuire et al., 2018). This subset of models also showed large reductions of near-surface permafrost area, averaging a 90% loss (12.7 ± 5.1x10^6 km^2) of permafrost area by 2300 for RCP8.5 and 29% loss (4.1 ± 0.6x10^6 km^2) for RCP4.5, with much of that long-term loss already occurring by 2100.

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 increase. 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; Kloster et al., 2012; Wotton et al., 2017). 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; Pastick et al., 2017). 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 increase the spread of fire on the landscape (Héon et al., 2014). A regional process-model study of Alaska projected annual median area burned during the 21st century to be 1.3-1.7 times higher compared to the historical average compared to the historical average (Pastick et al., 2017). Fire also appears to be expanding as a novel disturbance into tundra and forest-tundra boundary regions previously protected by a cool, moist climate (Jones et al., 2009; Hu et al., 2010; Hu et al., 2015) (*medium confidence*). Annual tundra area burned in Alaska is projected to double under RCP 6.0 from a historic rate of 270 km^2 per year to 500-610 km^2 per year over the 21st century (Hu et al., 2015). A statistical approach projected a fourfold increase in the 30-yr probability of fire occurrence in the forest-tundra boundary by 2100 (Young et al., 2017). In contrast to fire, there has not yet been a comprehensive circumpolar projection of how abrupt thaw rates may change in the future, but one component of abrupt thaw, change in abrupt thaw lake area, has been projected to increase to increase by 53% under RCP8.5 (Walter Anthony et al., 2018) above the 1.4 x 10^6 km^2 of small lakes and ponds that currently exist in the permafrost region (Muster et al., 2017). As a result, there is *low confidence* in the ability to assess the magnitude by which abrupt thaw across the entire landscape will affect regional permafrost, 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), with increased specific humidity due to enhanced evaporation (Laîné et al., 2014), and moisture flux convergence increases into the Arctic (Skific and Francis, 2013). 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). 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 CMIP5 model projections of increased discharge from Arctic watersheds (van Vliet et al., 2013; Gelfan et al., 2016; MacDonald et al., 2018). 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).

Lake ice phenology is sensitive to projected changes in surface temperature (Sharma et al., 2019). 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, with more extreme reductions for coastal regions (Brown and Duguay, 2011; Dibike et al., 2011; Prowse et al., 2011) (*medium confidence*). 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 complicate projections of these events (Prowse et al., 2010; Prowse et al., 2011).

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 CO_2 and methane acts as a feedback to accelerate global climate change. There is *high confidence* that the northern region acted as a 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 ecosystems 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*) (Webb et al., 2016) or the separation of upland and wetland ecosystem types, which was done in one synthesis but not the other. Moisture status is a primary control over ecosystem carbon sink/source strength with wetlands more often than not still acting as annual net carbon sinks even while methane is emitted (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 CO_2-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-C per year assuming the Alaska study region (1.6 x 10^6 km^2) could be scaled to the entire northern circumpolar permafrost region 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., 2018b) (*very high confidence*). Initial estimates were converging on a range of cumulative emissions from soils to the atmosphere by 2100, but recent studies have actually widened that range somewhat (Figure 3.11) (*medium confidence*). Expert assessment and laboratory 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 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 high emission climate warming trajectories, 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 modelling 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 (Kleinen and Brovkin, 2018) or all of these losses, at least during this century, by sequestering new carbon into plant biomass and increasing carbon inputs into the surface soil (McGuire et al., 2018). These future carbon emission levels would be a significant fraction of those projected from fossil fuels with implications for allowable carbon budgets that are consistent with limiting global warming, but will also depend on how vegetation responds (*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 currently contributing to increasing atmospheric concentrations. 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 similarly to other northern wetland ecosystems, or that increasing atmospheric methane concentrations in northern observation sites is derived from methane coming from midlatitudes (Sweeney et al., 2016). However, 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 ebullition (bubbling) flux to the atmosphere that is heterogeneous in time and space. Some new quantifications include: 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 based on methane concentrations in air and water has widened with more observations and now ranges from 3 Tg CH_4 yr^{-1} (Thornton et al., 2016) to 17 Tg CH_4 yr^{-1} (Shakhova et al., 2013). 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 dynamical 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). A model 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 2100 largely as a result of an increase in wetland extent (Zhang et al., 2017). However, 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 (McGuire et al., 2018). 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 *medium confidence* in future model projections of methane. Recent studies that addressed some of these landscape controls over future emissions projected increases in methane above the current levels on the order 10-60 Tg CH_4 yr^{-1} under RCP8.5 by 2100 (Schuur et al., 2013; Koven et al., 2015; Lawrence et al., 2015; Schneider von Deimling et al., 2015; Walter Anthony et al., 2018). These additional methane fluxes are projected to cause 40-70% of total permafrost-affected radiative forcing in this century even though methane emissions are much less than CO_2 by mass (Schneider von Deimling et al., 2015; Walter Anthony et al., 2018). As with total carbon emissions, there is *high confidence* that mitigation of anthropogenic methane sources could help to dampen the impact of increased methane emissions from the Arctic and boreal regions (Christensen et al., 2019).

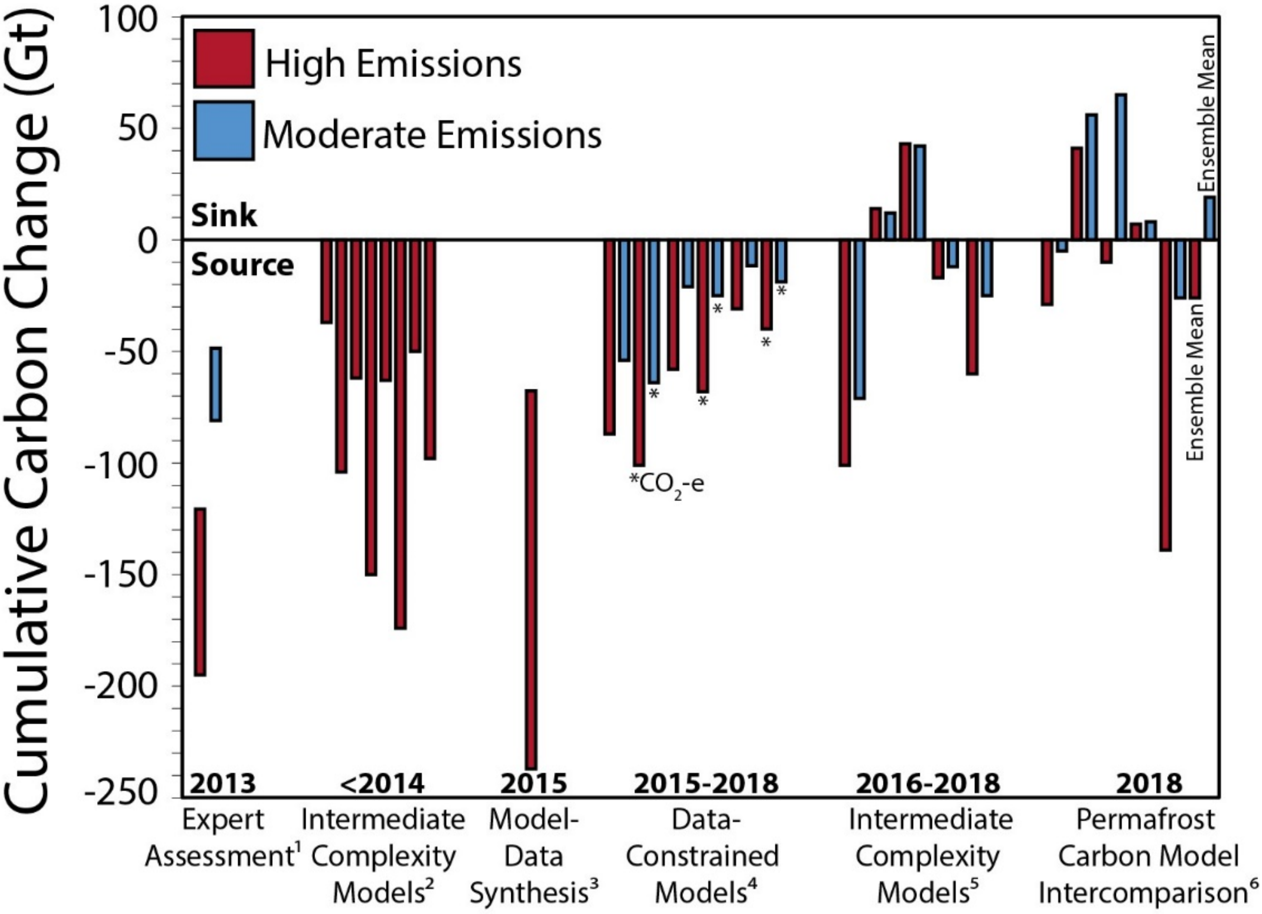


Figure 3.11: Estimates of cumulative net soil carbon pool change for the northern circumpolar permafrost region by 2100 following medium and high emission scenarios (e.g. RCP4.5 and RCP8.5 or equivalent). Cumulative carbon amounts are shown in Gigatons C (1 Gt C=1 billion metric tons), with source (negative values) indicating net carbon movement from soil to the atmosphere and sink (positive values) indicating the reverse. Some data-constrained models differentiated CO_2 and CH_4; bars show total carbon by weight, paired bar with * indicate CO_2-equivalent, which takes into account the global warming potential of CH_4. Ensemble mean bars refer to the model average for the Permafrost Carbon Model Intercomparison Project [5 models]. Bars that do not start at zero are in part informed by expert assessment and are shown as 95%CI ranges; all other bars represent model mean estimates. Data are from [1](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; Walter Anthony et al., 2018); [5](MacDougall and Knutti, 2016; Burke et al., 2017a; Kleinen and Brovkin, 2018); [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 provides a positive feedback to global temperatures (Flanner et al., 2011; Qu and Hall, 2014; Thackeray and Fletcher, 2016) (*high confidence*). Estimates of increases in global net solar energy flux due to snow cover loss range from 0.10 W m^{-2} to 0.22 W m^{-2} (± 50%; *medium confidence*) depending on dataset and time period (Flanner et al., 2011; Chen et al., 2015; Singh et al., 2015; Chen et al., 2016b). Sources of uncertainty include the range in observed spring snow cover extent trends (Hori et al., 2017) and the influence of clouds on shortwave feedbacks (Sedlar, 2018; Sledd and L'Ecuyer, 2019). Terrestrial snow changes also affect the longwave energy budget via altered surface emissivity (Huang et al., 2018). Climate model simulations show that changes in snow cover dominate land surface related positive feedbacks to atmospheric heating (Euskirchen et al., 2016), but regional variations in surface albedo are also influenced by vegetation (Loranty et al., 2014). There is evidence for positive sensitivity of surface temperatures to increased northern hemisphere boreal and tundra leaf area index, which contributes a positive feedback to warming (Forzieri et al., 2017).

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 across the 36–year satellite period (1982–2017) shows increasing above ground biomass (=greening) throughout a majority of the circumpolar Arctic (*high confidence*) (Xu et al., 2013a; Ju and Masek, 2016; Bhatt et al., 2017). Increasing greenness has been in some cases linked with shifts in plant species dominance away from graminoids (grasses and sedges) towards shrubs (*high confidence*) (Myers-Smith et al., 2015). Within the overall trend of greening, some tundra show declines in vegetation biomass (browning) (Bhatt et al., 2017).

The spatial variation in greening and browning trends in tundra are also not consistent over time (decadal scale) and can vary across landform/ecosystem types (Lara et al., 2018), suggesting interactions between the changing environment and the biological components of the system that control these trends. There is *high confidence* that increases in summer, spring, and winter temperatures lead to tundra greening, as well as increases in 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 thickness (via nutrient availability or changes in moisture), 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). 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*).

Projections of tundra vegetation distribution across the Arctic by 2050 in response to changing environmental conditions suggest that the areal extent of most tundra types will decrease by at least 50% (Pearson et al., 2013). Woody shrubs and trees are projected to expand to cover 24-52% of the current tundra region by 2050, or 12-33% if tree dispersal is restricted. Adding to this, 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) at faster rates than would occur by changing environmental conditions alone. In tundra regions, graminoid (grasses and sedges) 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 (Pastick et al., 2017).

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. While boreal forest may expand at the northern edge (Pearson et al., 2013), climate projections suggest that it could diminish at the southern edge and be replaced by lower biomass woodland/shrublands (Koven, 2013; Gauthier et al., 2015). Furthermore, 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 permafrost stability, hydrology, and vegetation. 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 (Johnstone et al., 2010). 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). This projected trend of increasing deciduous forest at the expense of evergreen forest is mirrored in Russian and Chinese boreal forests as well (Shakhova et al., 2013; Shuman et al., 2015; Wu et al., 2017).

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) (*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* changes. 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 snow depth reduce *Rangifer* survival (Mallory and Boyce, 2017). Changes in the timing of sea ice formation have direct effects on risks during *Rangifer* migration via inter-island movement and connection to the mainland (Poole et al., 2010). Summer warming is changing the composition of tundra plant communities, modifying the relationship between climate, forage, and *Rangifer* (Albon et al., 2017), which also impacts 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).

In northern Fennoscandia, there are approximately 600,000 semi-domesticated reindeer. Lichen rangelands are key to sustaining reindeer carrying capacity, with variable response 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). Later ice formation on waterbodies can impact herding activities (Turunen et al., 2016). Ice formation from rain-on-snow events is associated with population changes including cases of catastrophic mass starvation (Bartsch et al., 2010; Forbes et al., 2016), but there is no evidence of trends in rain-on-snow events (Cohen et al., 2015; Dolant et al., 2017).

Management of keystone species requires an understanding of pathogens and disease in the context of climate warming, but evidence of changing patterns across northern ecosystems (spanning terrestrial, aquatic, and marine environments) is hindered by an incomplete picture of pathogen diversity and distribution (Hoberg, 2013; Jenkins et al., 2013; Cook et al., 2017). Among ungulates, it is *virtually certain* that the emergence of disease attributed to nematode pathogens has accelerated since 2000 in the Canadian Arctic islands and Fennoscandia (Kutz et al., 2013; Hoberg and Brooks, 2015; Laaksonen et al., 2017; Kafle et al., 2018a). Discovery of the pathogenic bacterium *Erysipelothrix rhusiopathiae* has been linked to massive and widespread mortality among muskoxen from the Canadian Arctic Archipelago; loss of >50% of the population since 2010 may be attributable to disease interacting with extreme temperature events, although unequivocal links to climate have not been established (Kutz et al., 2015; Forde et al., 2016a; Forde et al., 2016b). Anthrax is projected to expand northward in response to warming, and resulted in substantial mortality events for reindeer on the Yamal Peninsula of Russia in 2016 with mobilization of bacteria possibly from a frozen reindeer carcass or melting permafrost (Walsh et al., 2018). In concert with climate forcing, pathogens are *very likely* responsible for increasing mortality in Arctic ungulates (muskox, caribou/reindeer) and alteration of transmission patterns in marine food chains, broadly threatening sustainability of subsistence and commercial hunting and fishing and safety of traditional foods for northern cultures at high latitudes (Jenkins et al., 2013; Kutz et al., 2014; Hoberg et al., 2017).

3.4.3.2.3 Freshwater

Climate-driven changes in seasonal ice and permafrost conditions influence water quality (*high confidence*). Shortened duration of freshwater ice cover (more light absorption, increased nutrient input) is expected to result in higher primary productivity (Hodgson and Smol, 2008; Vincent et al., 2011; Griffiths et al., 2017b) and may also encourage greater methane emissions from Arctic lakes (Greene et al., 2014; Tan and Zhuang, 2015). Thaw slumps, active layer detachments, and peat plateau collapse affect surface water connectivity (Connon et al., 2014) and enhance sediment, particulate and solute fluxes in river and stream networks (Kokelj et al., 2013). The transfer of enhanced nutrients from land to water (driven by active layer thickening and thermokarst processes; Abbott et al., 2015; Vonk et al., 2015) has been linked to heightened autotrophic productivity in freshwater ecosystems (Wrona et al., 2016). Still, there is *low confidence* in the influence of permafrost changes on dissolved organic carbon, because of competing mechanisms that influence carbon export. Permafrost thaw could contribute to the mobilization of previously frozen organic carbon (Abbott et al., 2014; Wickland et al., 2018; Walvoord et al., 2019) thereby enhancing both particulate and dissolved organic carbon export to aquatic systems. Increased delivery of this dissolved carbon from enhanced river discharge to the Arctic Ocean (Section 3.4.3.1.2) can exacerbate regionally-extreme aragonite undersaturation of shelf waters (Semiletov et al., 2016) driven by ocean uptake of anthropogenic CO_2 (Section 3.2.1.2.4). Conversely, reduced dissolved organic carbon export could accompany permafrost thaw as (1) water infiltrates deeper and has longer residence times for decomposition (Striegl et al., 2005) and (2) the proportion of groundwater (typically lower in dissolved organic carbon and higher in dissolved inorganic carbon than runoff) to total streamflow increases (Walvoord and Striegl, 2007). Increased thermokarst also has the potential to impact freshwater cycling of inorganic carbon (Zolkos et al., 2018).

Enhanced subsurface water fluxes resulting from permafrost degradation has consequences for inorganic natural and anthropogenic constituents. Emerging evidence suggests large natural stores of mercury (Schuster et al., 2018; St Pierre et al., 2018) and other trace elements in permafrost (Colombo et al., 2018) may be released upon thaw, thereby having effects (largely unknown at this point) on aquatic ecosystems. In parallel, increased development activity in the Arctic is *likely* to lead to enhanced local sources of anthropogenic chemicals of emerging Arctic concern, including siloxanes, parabens, flame retardants, and per- and polyfluoroalkyl substances (AMAP, 2017c). For legacy pollutants, there is *high confidence* that black carbon and persistent organic pollutants (e.g., hexachlorocyclohexanes, polycyclic aromatic hydrocarbons, and polychlorinated biphenyls) can be transferred downstream and affect water quality (Hodson, 2014). Lakes can become sinks of these contaminants, while floodplains can be contaminated (Sharma et al., 2015).

There is *high confidence* that habitat loss or change due to climate change impact Arctic fishes. Thinning ice on lakes and streams changes the overwintering habitat for aquatic fauna by impacting winter water volumes and dissolved oxygen levels (Leppi et al., 2016). Surface water loss, reduced surface water connectivity among aquatic habitats, and changes to the timing and magnitude of seasonal flows (Section 3.4.1.2) result in a direct loss of spawning, feeding, or rearing habitats (Poesch et al., 2016). Changes to permafrost landscapes have reduced freshwater habitat available for fishes and other aquatic biota, including aquatic invertebrates upon which the fish depend for food (Chin et al., 2016). Gullying deepens channels (Rowland et al., 2011; Liljedahl et al., 2016) that otherwise may connect lake habitats occupied by fishes. This can lead to the 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). There is regional evidence that migration timing has shifted earlier and winter egg incubation temperature has increased 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. Changes to the Arctic growing season (Xu et al., 2013a) increase the risk of drying of surface water habitats and pose 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 the impact on lower trophic levels within food webs is not clearly understood.

[START BOX 3.4 HERE]

Box 3.4: Impacts and Risks for Polar Biodiversity from Range Shifts and Species Invasions Related to Climate Change

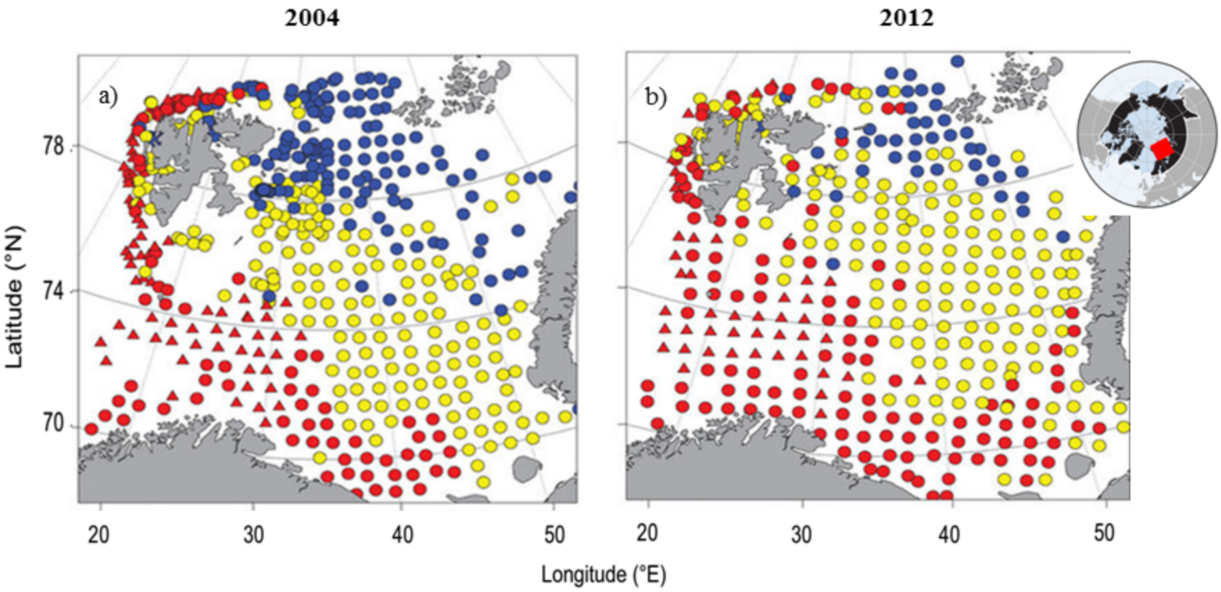
In polar regions climate-induced changes in terrestrial, ocean and sea ice environments, together with human introduction of non-native species, have expanded the range of some temperate species and contracted the range of some polar fish and ice associated species (Section 3.2.3.2; Duffy et al., 2017) (*high confidence* for detection*, medium confidence* for attribution). In some cases, spatial shifts in distribution have also been influenced by fluctuations in population abundance linked to climate-induced impacts on reproductive success (Section 3.2.3). These changes have the potential to alter biodiversity in polar marine and terrestrial ecosystems (Frenot et al., 2005; Frederiksen, 2017; McCarthy et al., 2019) (*medium confidence*).

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) (*high confidence*). This is further complicated by the northward expansion of the summer ranges of a variety of temperate whale species, documented recently in both the Pacific and Atlantic sides of the Arctic (Brower et al., 2017; Storrie et al., 2018) and increasing pressure from anthropogenic activities. Also, over the recent decade a northward shift in benthic species, with subsequent changes in community composition has been detected in both the northern Bering Sea (Grebmeier, 2012), off Western Greenland (Renaud et al., 2015), and the Barents Sea (Kortsch et al., 2012) (*medium confidence*). At the same time as these northward expansions or shifts, a number of populations of species as different as polar bear and Arctic char show range contraction or population declines (Winfield et al., 2010; Bromaghin et al., 2015; Laidre et al., 2018).

In the Arctic a number of fish species have changed their spatial distribution substantially over the recent decades (*high confidence*). The most pronounced recent range expansion into the Arctic of all may be that of the summer feeding distribution of the temperate Atlantic mackerel (*Scomber scombrus*) in the Nordic Seas. From 1997 to 2016 the total area occupied by this large stock expanded from 0.4 to 2.5 million km^2 and the centre-of-gravity of distribution shifted westward by 1650 km and northward by 400 km (Olafsdottir et al., 2019), far into Icelandic and Greenland waters and even up to Svalbard (Berge et al., 2015; Jansen et al., 2016; Nøttestad et al., 2016). This range expansion was linked both to a pronounced increase in stock size and warming of the ocean (Berge et al., 2015; Olafsdottir et al., 2019) (*high confidence*). Under RCP4.5 and RCP8.5 further range expansions of mackerel are projected in Greenland waters (Jansen et al., 2016) (*medium confidence*). However, further northwards expansion of planktivorous species may generally be restricted by them not being adapted to lack of primary production during winter (Sundby et al., 2016). Range shifts have also been observed in the Bering Sea since 1993 with warm bottom temperatures being associated with range contractions of Arctic species, and range expansions of sub-arctic species, with responses dependent on species specific vulnerably (Alabia et al., 2018; Stevenson and Lauth, 2018).

In the Barents Sea, major expansions in distribution over the recent years to decades have been well documented for both individual species and whole biological communities (*high confidence*). New information strengthens findings reported in WGII AR5 of ecologically- and commercially-important fish stocks having extended their habitats markedly to the north and east, concomitant to increased sea temperature and retreating sea ice. This includes capelin (Ingvaldsen and Gjøsæter, 2013), Atlantic cod (Kjesbu et al., 2014), and haddock (Landa et al., 2014). Of even greater importance is novel evidence of distinct distributional changes at the community level (Fossheim et al., 2015; Kortsch et al., 2015; Frainer et al., 2017) (Box 3.4 Figure 1). Until recently, the northern Barents Sea was dominated by small-sized, slow-growing fish species with specialized diets, mostly living in close association with the sea floor. Simultaneous with rising sea temperatures and retreating sea ice, these Arctic fishes are being replaced by boreal, fast-growing, large-bodied generalist fish moving in from the south. These large, migratory predators take advantage of increased production while the Arctic fish species suffer from higher competition and predation and are retracting northwards and eastwards. Consequently, climate change is inducing structural change over large spatial scales, leading to a borealization ('Atlantification') of the European Arctic biological communities (Fossheim et al., 2015; Kortsch et al., 2015; Frainer et al., 2017) (*medium confidence*).

There is evidence based on population genetics that the ecosystem off Northeast Greenland is also *likely* to become populated by a larger proportion of boreal species with ocean warming. Andrews et al. (2019) show that Atlantic cod, beaked redfish (*Sebastes mentella*), and deep-sea shrimp (*Pandalus borealis*) recently found on the Northeast Greenland shelf originate from the quite distant Barents Sea, and suggested that pelagic offspring were dispersed via advection across the Fram Strait.



Box 3.4, Figure 1: Spatial distribution of fish communities identified at bottom trawl stations in the Barents Sea (north of northern Norway and Russia, position indicated by red box in small globe) in (a) 2004 and (b) 2012. Atlantic (red), Arctic (blue) and Central communities (yellow). Circles: shallow sub-communities, triangles: deep sub-communities. Modified from Fossheim et al. (2015).

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 future expansions of fish populations (*medium confidence*). The limited available information on marine fish from other Arctic shelf regions reveals a latitudinal cline in the abundance of commercially-harvestable fish species. For instance, there is evidence of latitudinal partitioning between the four dominant mid-water species (Polar cod, saffron cod [*Eleginus gracilis*], capelin, and Pacific herring, [*Clupea pallasii*]) in the Chuckchi and Northern Bering Sea, with Polar cod being most abundant to the north (De Robertis et al., 2017). These latitudinal gradients suggest that future range expansions of fish populations will continue to be governed by a combination of physical factors affecting overwintering success and the availability, quality and quantity of prey (*medium confidence*).

In Antarctic marine systems, there is evidence of recent climate-related range shifts in the southwest Atlantic and West Antarctic Peninsula for penguin species (*Pygoscelis papua* and *P. antarctica*) and for Antarctic krill (*Euphausia superba*), but mesozooplankton communities do not appear to have changed or shifted in response to ocean warming (Section 3.2.3.2). Recent evidence suggests that the Antarctic Circumpolar Current and its associated fronts and thermal gradients may be more permeable to biological dispersal than previously thought, with storm-forced surface waves and ocean eddies enhancing oceanographic connectivity for drift particles in surface layers of the Southern Ocean (Fraser et al., 2017; Fraser et al., 2018) (*low confidence*), but it is unclear whether this will be an increasingly-important pathway under climate change. Greater ship activity in the Southern Ocean may also present a risk for increasing introduction of non-native marine species, with the potential for these species to become invasive with changing environmental conditions (McCarthy et al., 2019). Current evidence of invasions by shell-crushing crabs on the Antarctic continental slope and shelf remains equivocal (Griffiths et al., 2013; Aronson et al., 2015; Smith et al., 2017d).

On Arctic land, northward range expansions have been recorded in species from all major taxon groups based both on scientific studies and local observations (*high confidence*) (CAFF, 2013b; AMAP, 2017a; AMAP, 2017b; AMAP, 2018). The most recent examples of terrestrial vertebrates expanding northwards include a whole range of mammals in Yakutia, Russia (Safronov, 2016), moose (*Alces alces*) into the Arctic region 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 also spreading north (CAFF, 2013b; Taylor et al., 2015; Forde et al., 2016b; Burke et al., 2017b; Kafle et al., 2018a). A widespread change is tundra greening, which in some cases is linked to shifting plant dominance within Arctic plant communities, in particular an increase in woody shrub biomass as conditions become more favorable for them (Myers-Smith et al., 2015; Bhatt et al., 2017).

Expansion of subarctic terrestrial species and biological communities into the Arctic and displacing native species is considered a major threat, since unique Arctic species may be less competitive than encroaching subarctic species favoured by changing climatic conditions (CAFF, 2013b). Similar displacements may take place within zones of the Arctic when Low- and Mid-Arctic species expand northward. Here, the most vulnerable species and communities may be in the species-poor, but unique, northernmost sub-zone of the Arctic because species cannot migrate northward as southern species encroach (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 poleward and that there is nowhere further north on land to go for terrestrial biota at the northern coast. The expected overall result of these shifts and limits will be a loss of biodiversity (CAFF, 2013b; CAFF, 2013a; 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., 2015) 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 1–2°C in the northernmost subzone may allow the establishment of woody dwarf shrubs, sedges and other species into bare soil areas that may radically change its appearance and ecological functions (Walker et al., 2015; Myers-Smith et al., 2019) (*medium confidence*).

Range expansions also include the threat from alien species brought in by humans to become invasive and outcompete native species. Relatively few invasive alien species are presently well established in the Arctic, but many are thriving in the subarctic and may expand as a result of climate change (CAFF, 2013b; CAFF, 2013a). Examples of this include: American mink (*Neovison vison*) and Nootka lupin (*Lupinus nootkatensis*) in Arctic western Eurasia, Greenland and Iceland that are already causing severe problems to native fauna and flora (CAFF and PAME, 2017).

Alien species are a major driver of terrestrial biodiversity change also in the Antarctic region (Frenot et al., 2005; Chown et al., 2012; McClelland et al., 2017). The Protocol on Environmental Protection to the Antarctic Treaty restricts the introduction of non-native species to Antarctica as do the management authorities of sub-Antarctic islands (De Villiers et al., 2006). Despite this, alien species and their propagules continue to be introduced to the Antarctic continent and sub-Antarctic islands (Hughes et al., 2015). To date, 14 non-native terrestrial species have colonised the Antarctic Treaty area (excluding subantarctic islands) (Hughes et al., 2015), while the number in the subantarctic is much higher (on the order of 200 species) (Frenot et al., 2005). Species distribution models for terrestrial invasive species indicate that climate does not currently constitute a barrier for the establishment of invasive species on all subantarctic islands, and that the Antarctic Peninsula region will be the most vulnerable location on the Antarctic continent to invasive species establishment 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). Along with growing number of visitors, this is expected to increase in the establishment probability of terrestrial alien species (Chown et al., 2012; Hughes et al., 2015) (*medium confidence*).

[END BOX 3.4 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 for indigenous communities (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 Food and water 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).

There is *high confidence* in indicators that food insecurity risks are on the rise for Indigenous Arctic peoples. 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. Food systems in northern communities are intertwined with northern ecosystems because of subsistence hunting, fishing, and gathering activities. Environmental changes to animal habitat, population sizes, and movement mean that culturally-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 impacts negatively 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). 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). Limiting exposures to zoonotic, foodborne, and waterborne pathogens (Section 3.4.3.2.2) depends on accurate and comprehensive data on species diversity, biology and distribution, and pathways for invasion (Hoberg and Brooks, 2015; Kafle et al., 2018b).

There is *high confidence* that changes to travel conditions impact food security through access to hunting grounds. Shorter snow cover duration (Section 3.4.1.1), and changes to snow conditions (such as density) make travel more difficult and dangerous (Laidler, 2012; Ford et al., 2019). 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 and landscape instability (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; Dodd et al., 2018).

There is *high confidence* that both risks and opportunities arise for coastal communities with changing sea ice and open water conditions. 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). 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). 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).

Many northern communities rely on ponds, streams, and lakes for drinking water (Cochran et al., 2013; Goldhar et al., 2013; Nymand and Fondahl, 2014; Daley et al., 2015; Dudley et al., 2015; Masina et al., 2019), 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, the risks of which are increased by warming ground and water temperatures (Cozzetto et al., 2013; Goldhar et al., 2013; Dudley et al., 2015; Masina et al., 2019). Icebergs or old multi-year ice are important sources of drinking water for some coastal communities, so reduced accessibility to stable sea ice conditions affects local water security. Small remote communities have limited capacity to respond quickly to water supply threats, which amplifies vulnerabilities to 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). There is *high confidence* that daily life is influenced by changes 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 because younger generations do not have the same level of experience or confidence with traditional indicators (Ford, 2012; Parlee and Furgal, 2012; Eicken et al., 2014; Pearce et al., 2015). Climate-driven changes undermine confidence in indigenous knowledge holders in regards to traditional indicators used for safe travel and navigation (Parlee and Furgal, 2012; Golovnev, 2017; Ford et al., 2019).

Economics

The Arctic mixed economy is characterized by a combination of subsistence activities, and employment and cash income. There is *low confidence* about the extent and nature of impact of climate change on local subsistence activities and economic opportunities across the Arctic (e.g., hunting, fishing, resource extraction, tourism and transportation; see Section 3.2.4) because of high variability between communities (Harder and Wenzel, 2012; Cochran et al., 2013; Clark et al., 2016b; Fall, 2016; Ford et al., 2016; Lavrillier et al., 2016). 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 (Sections 3.2.4.2, 3.2.4.3) (Ford et al., 2012; Huskey et al., 2014; Overland et al., 2017). This has important implications for economic development, particularly in relation to local employment opportunities but also raises concerns of detrimental impacts on animals, habitat, and subsistence activities (Cochran et al., 2013; Inuit Circumpolar Council, 2015).

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; Ostapchuk et al., 2015). Climate change consequences in polar regions (Sections 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 (Durkalec et al., 2014; Durkalec et al., 2015; Clark et al., 2016b; Driscoll et al., 2016).

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). Mercury presently stored in permafrost (Schuster et al., 2018) has potential to accumulate in aquatic ecosystems.

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 (Section 3.4.3.2.3) or potential waterborne diseases (Goldhar et al., 2014; Daley et al., 2015). The potential for infectious gastrointestinal disease is not well understood, and there are concerns in relation to the safety of storage containers of raw water in addition to the quality of the source water itself (Goldhar et al., 2014; Wright et al., 2017; Masina et al., 2019).

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; Ostapchuk 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 (Ostapchuk 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 health, as well as spiritual and social vitality according to relationships between sea ice use, culture, knowledge, and autonomy (Cunsolo Willox et al., 2013a; Cunsolo Willox et al., 2013b; 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, 2017d). The observed changes in the ground thermal regime (Romanovsky et al., 2010; Romanovsky et al., 2017; Biskaborn et al., 2019) threaten the structural stability and functional capacities of infrastructure, in particular that which is located on ice-rich frozen ground. Extensive summaries of construction damages along with adaptation and mitigation strategies are available (Larsen et al., 2014; Dore et al., 2016; AMAP, 2017d; Pendakur, 2017; Shiklomanov et al., 2017a; Shiklomanov et al., 2017b; Vincent et al., 2017).

Projections of climate and permafrost suggest that a wide range current infrastructure will be impacted by changing conditions (*medium confidence*). A circumpolar study found that approximately 70% of 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 under RCP4.5 (Hjort et al., 2018). 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. 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.

In a regional study of the state of Alaska, cumulative expenses projected for climate-related damage to public 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% to USD4.2 billion for the same time frame under RCP4.5, indicating that reducing greenhouse gas emissions globally could lessen damages (Figure 3.13). In a related study that included these costs and others, as well as positive gains from climate change in terms of a reduction in heating costs attributable to warmer winter, annual net costs were still USD340–$700 million, or 0.6%–1.3% of Alaska's GDP, suggesting that climate change costs will outweigh positive benefits, at least for this region (Berman and Schmidt, 2019).

Winter roads (snow covered ground and frozen lakes) are distinct from the infrastructure considered earlier, but have a strong influence on the reliability and costs of transportation in some remote northern communities and industrial development sites (Parlee and Furgal, 2012; Huskey et al., 2014; Overland et al., 2017). For these 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). There have been recent instances of severely curtailed ice road shipping seasons due to unusually warm conditions in the early winter (Sturm et al., 2017). While the impact of human effort on the maintenance of winter roads is difficult to quantify, a reduction in the operational time window due to winter warming is projected (Mullan et al., 2017).

3.5 Human Responses to Climate Change in Polar Regions

3.5.1 The Polar Context for Responding

Human responses to climate change in the Arctic and Antarctica are shaped by their unique physical, ecological, social, cultural, and political conditions. Extreme climatic conditions, remoteness from densely populated regions, limited human mobility, short seasons of biological productivity, high costs in monitoring and research, sovereignty claims to lands and waters by southern-based governments, a rich diversity of indigenous cultures, and institutional arrangements that in some cases recognize indigenous rights and support regional and international cooperation in governance are among the many factors that impede and or facilitate adaptation.

The social and cultural differences are an especially noteworthy factor in assessing polar responses. Approximately 4 million people currently reside in the Arctic with about three quarters residing in urban areas, and approximately 10% being Indigenous (AHDR, 2014). Regions of the Arctic differ widely in population, ranging from 94% of Iceland's population living in urban environments to 68% of Nunavut, Canada's population living in rural areas. And while there has been a general movement to greater urbanization in the Arctic (AHDR, 2014), that trend is not true for all regions (Heleniak, 2014). About 4400 people reside in Antarctic in the summer and about 1100 in the winter, predominantly based at research stations of which approximately 40 are occupied year-round (The World Factbook, 2016).

For most Arctic Indigenous peoples, human responses to climate change are viewed as a matter of cultural survival (Greaves, 2016) (Cross-Chapter Box 3 in Chapter 1). However, Indigenous people are not homogenous in their perspectives. 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).* Geopolitical complexities also confound responses.

Together these conditions make for complexity and uncertainty in human decision making, be it at the household and community levels to the international level. Adding to uncertainty in human choice related to climate change is the interaction of climate with other forces for change, such as globalization and land and sea-use change. These interactions necessitate that responses to climate change consider cumulative effects as well as context-specific pathways for building resilience (Nymand and Fondahl, 2014; ARR, 2016).

3.5.2 Responses of Human Sectors

The sections below assess human responses to climate change in polar regions by examining various sectors of human-environment activity (i.e. social-ecological subsystems), reviewing their respective systems of governance related to climate change, and considering possible resilience pathways. Table 3.4 summarizes the consequences, interacting drivers, responses, and assets for responding to climate change by social-ecological subsystems (i.e., sectors) of Arctic and Antarctic regions. An area of response not elaborated in this assessment is geo-engineered sea ice remediation to support local-to-regional ecosystem restoration and which may also affect climate via albedo changes. There is an emerging body of literature on this topic (e.g., Berdahl et al., 2014; Desch et al., 2017; Field et al., 2018), which at present is too limited to allow assessing dimensions of feasibility, benefits and risks, and governance.

3.5.2.1 Commercial Fisheries

Responses addressing changes in the abundance and distribution of fish resources (Section 3.2.4.1) differ by region. In some polar regions, strategies of adaptive governance, biodiversity conservation, scenario planning, and the precautionary approach are in use (NPFMC, 2018). Further development of coordinated monitoring programs (Cahalan et al., 2014; Ganz et al., 2018), 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 (Busch et al., 2016; NPFMC, 2019; see Section 3.5.4). Scenario planning, adaptive management, and similar efforts will contribute to the resilience and conservation of these social-ecological 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 nautical mile Exclusive Economic Zones (EEZs) in the High Arctic and manage their resources within their own regulatory measures. A review of future harvest in the European Arctic (Haug et al., 2017) points towards high probability of increased northern movement of several commercial fish species (Section 3.3.3.1, Box 3.4), but only to the shelf slope for the demersal species. This shift suggests increased northern fishing activity, but within the EEZs and present management regimes (Haug et al., 2017) (*medium confidence*).

In 2009, a new Marine Resources Act entered into force for Norway's EEZ. This act applies to all wild living marine resources, and states that its purpose is to ensure sustainable and economically-profitable management of 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). 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*). A scenario-based approach to identify management strategies that are effective under changing climate conditions is being explored for the Barents Sea (Planque et al., 2019).

In the U.S. Arctic an adaptive management approach has been introduced that utilises 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 includes 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; Holsman et al., 2019). Despite this complex ecosystem-based approach to fisheries management, it may not be possible to prevent projected declines of some high-value species at high rates of global warming (Ianelli et al., 2016).

In the US portion of the Chukchi and Beaufort Seas EEZ, fishing is prohibited until sufficient information is obtained to sustainably manage the resource (Wilson and Ormseth, 2009). In the Canadian sector of the Beaufort Sea, commercial fisheries are currently only small scale and locally operated. However, with decreasing ice cover and potential interest in expanding fisheries, the Inuvialuit subsistence fishers of the western Canadian Arctic, developed 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 nation's EEZ. Recent actions of the international community show 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 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. The Arctic 5 and several other nations subsequently agreed to a treaty that imposed a 16-year moratorium on commercial fishing in the CAO. Several other agreements have adopted the same approach, including the Central Arctic Ocean Fisheries (CAOF) Agreement.

The Commission for the Conservation of Antarctic Marine Living Resources (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.

Commercial fisheries management responses to climate change impacts in the Southern Ocean may need to address the displacement of fishing effort due to poleward shifts in species distribution (Pecl et al., 2017) (Box 3.4) (*low confidence*). Fisheries in the Southern Ocean are relatively mobile and are potentially able to respond to range shifts in target species, which is in contrast to small-scale coastal fisheries in other regions. Management responses will also need to adapt to the effects of future changes in sea ice extent and duration on the spatial distribution of fishing operations (ATCM, 2017; Jabour, 2017) (Section 3.2.4).

3.5.2.2 Arctic Subsistence Systems

Subsistence users have responded to climate change by adapting their wildfood production systems and engaging in the climate policy processes at multiple levels of governance. The limitations of many formal institutions, however, suggest that in order to achieve greater resilience of subsistence systems with climate change, transformations in governance are needed to provide greater power sharing, including more resources for engaging in climate change studies and regional-to-national policy making (See 3.2.4.1.1, 3.4.3.2.2, 3.4.3.3.1, 3.4.3.3.2, 3.4.3.3.3, 3.5.3).

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

The impacts of climate change have also required adaptation to the non-harvesting aspects of wildfood production, such as an abandonment of traditional food storage and drying practices (e.g., ice cellars) and an increased use of household and community freezers (AMAP, 2017a). In 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 could 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 stronger links of local-to-regional level management with national- to international-level agreements necessitate 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 could be an efficient tool to achieve consensus on population goals, including international cooperation and agreements regarding migratory species shared between more countries (Kocho-Schellenberg and Berkes, 2014) (Section 3.5.4.3). While there has been involvement of subsistence users in monitoring and research on climate change (Section 3.5.4.1.1), resource management regimes that regulate harvesting are largely dictated by science-based paradigms that give limited legitimacy to the knowledge and suggested preferences of subsistence users (Section 3.5.4.2, Cross-Chapter Box 4 in Chapter 1).

The social costs and social learning associated with responding to climate change are often related. 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) (Forbes et al., 2015). In some cases, co-management has given communities a greater voice in decision making, but when ineffective, these arrangements can perpetuate dominant paradigms of resource management (AMAP, 2018). The perceived risks of climate change can at the same time reinforce cultural identify and motivate greater political involvement, which in turn, gives indigenous leaders experience as agents of change in policy making. Penn et al. (2016) pointed to these conflicting forces, arguing the need for a greater focus on community capacity and cumulative effects.

Greater involvement of indigenous subsistence users in Canada occurs at the national and regional levels through the structures and provisions of indigenous settlement agreements (e.g., 1993 Nunavut Land Claims Agreement, 1984 Inuvialuit Final Agreement), fish and wildlife co-management agreements (e.g., Porcupine Caribou Management Agreement of 1986), and through various boundary organizations (e.g, CircumArctric Rangifer Monitoring and Assessment Network). Home rule in Greenland, established in 1979, gives the Naalakkersuisut (government of Greenland) authority on most domestic matters of governance.

Indigenous leaders are responding to the risks of climate change by engaging in political processes at multiple levels and through different venues. However, indigenous involvement in IPCC assessments remains limited (Ford et al., 2016). At the United Nations Framework Convention on Climate Change (UNFCCC), the discursive space for incorporating perspectives of Indigenous peoples on climate change adaptation has expanded since 2010, which is reflected in texts and engagement with most activity areas (Ford et al., 2015) and by the establishment of the Local Communities and Indigenous Peoples Platform Facilitative Working Group in December 2018. Aleut International Association, Arctic Athabaskan Council, Gwich'in Council International, Inuit Circumpolar Council, 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 its working groups and partake also at the political level (Section 3.5.3.2.1).

3.5.2.3 Arctic Reindeer Herding

Herders’ responses to climate change have varied by region and respective herding practices, and in some cases are 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 the movements of caribou herds (wild reindeer), both of which are partially driven by climate. For Nenets of the Yamal, Russia, resilience in herding has been facilitated through herders’ own agency and, to some extent, the willingness of the gas industry 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 migration routes or take 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 (Section 3.4.3.2.2) (Forbes et al., 2016). In all these regions, restrictions affecting the movement of reindeer to pastures are expected to negatively interact with the effects of climate, and affect the future sustainability of herding systems (*high confidence*).

3.5.2.4 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 (Section 3.2.4.2). Polar-class expedition cruise vessels are now, for the first time, being purposefully built for recreational Arctic sea travel. The anticipated near- and long-term growth of Arctic tourism, especially with small vessels (yachts) (Johnston et al., 2017), points to a deficiency in current regulations and policies to address human safety, environmental risks, and culture al impacts. Industry growth also points to the need for operators, governments, destination communities, and others to identify and evaluate adaptation strategies, such as disaster relief management plans, updated navigation technologies for vessels, codes of conduct for visitors, and improved maps (Pizzolato et al., 2016) and to respond to perceptions of tourism by residents of local destinations (Kaján, 2014; Stokke and Haukeland, 2017). Efforts were initiated with stakeholders in Arctic Canada to identify strategies that would lower risks (Pizzolato et al., 2016); a next step to lower risks and build resilience is to further develop those strategies (AMAP, 2017a; AMAP, 2017b; AMAP, 2018). Opportunities for tourism vessels in the Arctic to contribute to international research activities ('ships of opportunity') may improve sovereignty claims in some regions, contribute to science, and enhance education of the public (Stewart et al., 2013; Arctic Council, 2015a; Stewart et al., 2015; de la Barre et al., 2016).

Tourism activities in the Antarctic are conducted in accordance with the Protocol on Environmental Protection to the Antarctic Treaty, which establishes general environmental principles, environmental assessment requirements, a scheme of establishing protected areas, and restrictions on waste disposal. Site-specific management tools are in place. While there are varying views amongst Antarctic Treaty Parties on the best management regulations for Antarctic tourism, these Parties continue to work to manage tourism activity, including growth in numbers of visitors. In addition to the Protocol, mandatory measures have been agreed to manage aspects of tourism activity. Industry self-regulation supplements these requirements, coordinated by the International Association of Antarctica Tour Operators (IAATO), which has worked with Antarctic Treaty Consultative Parties to manage changes in operations and their impact on ice-free areas (ATCM, 2016).

3.5.2.5 Arctic Non-Renewable Extractive Industries

Climate change has resulted a limited response by non-renewable resource extraction industries and agencies in the Arctic to changes in sea ice, thawing permafrost, spring run offs, and resultant timing of exploration, construction and use of ice roads, and infrastructure design (AHDR, 2014). In some regions, climate change has offered new development opportunities, although off-shore prospects remaining cost prohibitive given current world markets (Petrick et al., 2017). (In the area covered by the Antarctic Treaty, exploitation of mineral resources is prohibited by the Protocol on Environmental Protection to the Antarctic Treaty.)

Climate change in some Arctic regions is facilitating easier access to natural resources (Section 3.5.2.3), which may generate financial capital for Arctic residents and their governments, including Indigenous peoples but also greater exposure to risks such as oil spills and increases in noise. 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 mineral development commenced in Greenland, its home rule government developed environmental impact assessment protocols to provide for improved public participation (Forbes et al., 2015). Indigenous peoples are considered as non-state actors and in many, but not all cases, promote environmental protection in support of the sustainability of their traditional livelihoods. This protection is at times in opposition to the industrial 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 state-dominated companies (Young, 2016). Among the non-state actors, new networks and economic forums have been established (Wehrmann, 2016). One example is the Arctic Economic Council, created by the Arctic Council during 2013-15 as an independent organization that facilitates Arctic business-to-business activities and supports economic development.

Several regional governments are assessing the long-term viability of ice roads, historically used for accessing mineral development sites, as well as some Arctic human settlements. In Northwest Territories, Canada, several ice roads are being replaced with all-season roads, with other replacements proposed. Assessing future conditions is key for planning and initiating new projects (Hori et al., 2018; Kiani et al., 2018) but is often constrained by uncertainties of available climate models (Mullan et al., 2017).

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. 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 (Kirkfeldt et al., 2016).

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-responsible development. Better understanding of environmental change is also important in ensuring continued oil field operations with protection of natural resources. Improved forecasting of short-term conditions (i.e., snow, soil temps, spring run offs) could allow management agencies to respond to conditions more proactively, and give industry more time to plan winter mobilization, such as construction of ice roads (*low confidence*).

3.5.2.6 Infrastructure

Reducing and avoiding the impacts of climate change on infrastructure will require special attention to engineering, land-use planning, maintenance operations, local culture, and private and public budgeting (AMAP, 2017a; AMAP, 2017b; AMAP, 2018). In some cases, relocation of human settlements will be required, necessitating more formal methods of assessing relocation needs and identifying sources of funding to support relocations (Cross-Chapter Box 9) (*high confidence*).

A discussion of the relocation of Alaska's coastal villages is found in Cross-Chapter Box 9. Alaskan coastal communities are not the only settlements potentially requiring relocation. Subsidence due to thawing permafrost and river and delta erosion makes other rural communities of Alaska and Russia vulnerable, potentially requiring relocation in the future (Bronen, 2015; Romero Manrique et al., 2018). These situations raise issues of environmental justice and human rights (Bronen, 2017), and illustrate the limits of incremental adaptation when transformation change is needed (Kates et al., 2012). In other cases, cultural resources in the form of historic infrastructure are being threatened and require mitigation (Radosavljevic et al., 2015). Responsibility for funding has been a key issue in the relocation process (Iverson, 2013) as well as the overall role of government and local communities in relocation planning (Marino, 2012; Romero Manrique et al., 2018). The Alaska Denali Commission, an independent federal agency designed to provide critical utilities, infrastructure and economic support throughout Alaska, is now serving as the lead coordinating organization for Alaska village relocations and managing federal funding allocations. Several efforts have also been undertaken to provide assessment frameworks and protocols for settlement relocation as an adaptive resource (Bronen, 2015; Ristroph, 2017).

While there has been discussion of future 'climigration' in rural Alaska (Bronen and Chapin, 2013; Matthews and Potts, 2018), a study of Alaska rural villages threated by climate change showed no outmigration response (Hamilton et al., 2016). Several factors explain the lack of outmigration, including an unwillingness to move, attachment to place, people's inability to relocate, the effectiveness of alternative ways of achieving acceptable outcomes, and methods of buffering through subsidies (Huntington et al., 2018) (*medium confidence*).

The current pan-Arctic trend of urbanization (AHDR, 2014), suggests that climate change responses related to infrastructure in towns and cities of the North will require significant adaptation in designs and increases in spending (Streletskiy et al., 2012). These costs do not include costs related to flooding and other stressors associated with warming or additional costs of commercial and industrial operations. Engineers in countries with permafrost are actively working to adapt the design of structures to degrading permafrost conditions

(Dore et al., 2016) and the effects of a warming climate, for example the Cold Climate Housing Research Center of Alaska.

An analysis of the costs of total damages from climate change to public infrastructure in Alaska show the financial benefits of proactive adaptation (Melvin et al., 2017) (Figure 3.13). In addition to global carbon emission mitigation, hardening and redesigning of infrastructure can reduce costs of future climate-related impacts. For example, retrofitting and redesign infrastructure in order to handle increased precipitation and warmer temperatures can reduce climate-related costs by 50%, from $5.5 billion to $2.9 billion under RCP8.5 by 2100. The cost savings of retrofitting and redesigning infrastructure is even higher than the savings from carbon mitigation, where impact costs are estimated at $4.2 billion under RCP4.5 by 2100. Engineering adaptation provide proportionally similar cost savings no matter which emission scenario was used.

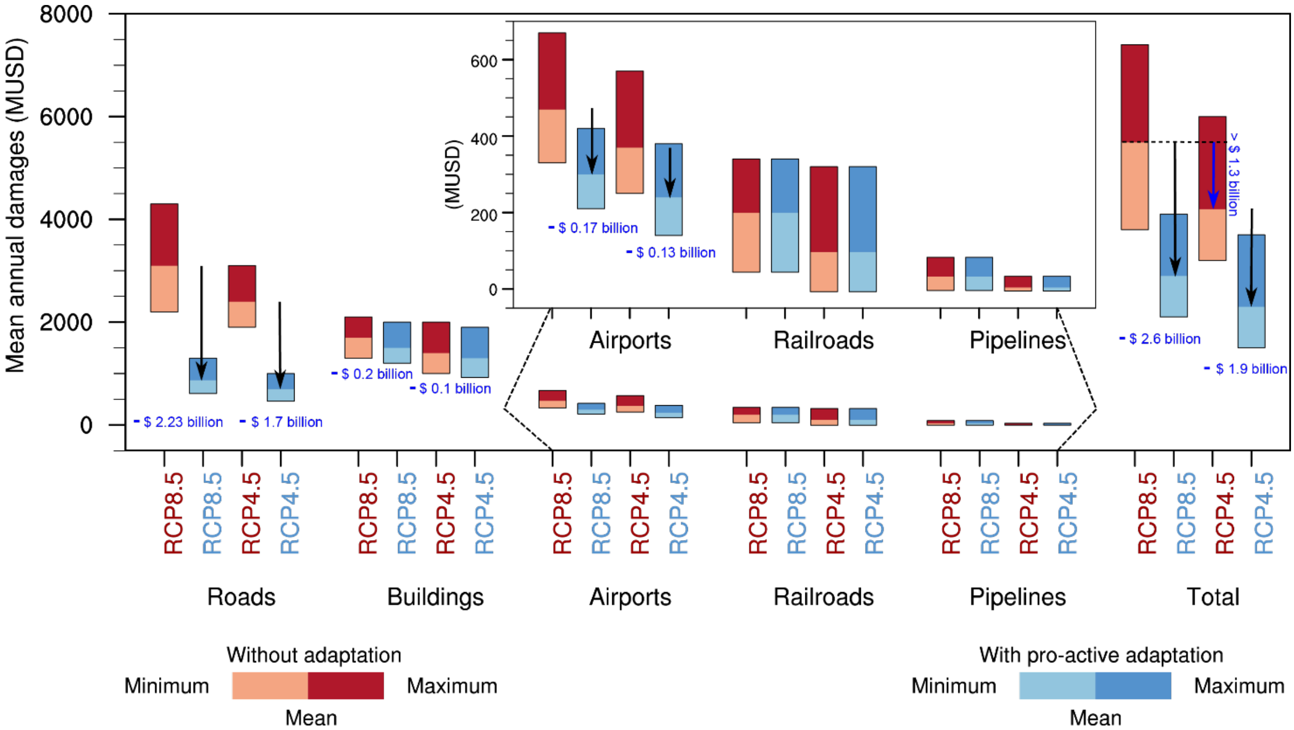


Figure 3.12: Changes in public infrastructure damage costs in cumulative $USD by 2100 in Alaska under different emission scenarios. The inset showing airports, railroads, and pipelines has a different in scale than roads, buildings, and the total. Dark shades represent climate-related costs of impact with no engineering adaptation measures, whereas light shades represent the cost savings after engineering adaptation (figure modified from Melvin et al., 2017).

3.5.2.7 Marine Transportation

Increases in Arctic marine transportation create impacts and risks for ecosystems and people, such as an increased likelihood of accidents, the introduction of invasive species, oil spills, waste discharges, detrimental impacts on animals, habitat, and subsistence activities (Sections 3.2.4.3, 3.4.3.3.2). There has been a rise in geopolitical debate regarding national- and international-level regulations and policies, and maritime infrastructure to support Arctic shipping development (Heininen and Finger, 2017; AMAP, 2018; Drewniak et al., 2018; Nilsson and Christensen, 2019). Without further action leading to adequate implementation of well-developed management plans and region-specific regulations, anticipated future increases in Arctic shipping will pose a greater risk to people and ecosystems (*high confidence*).

The International Maritime Organization has responsibility for the safety and security of shipping and the prevention of marine and atmospheric pollution by ships, including in the Arctic and Antarctic. There are a number of mechanisms standardizing regulation and governance, such as the International Convention for the Prevention of Pollution from Ships; the International Convention for the Safety of Life at Sea; the International Convention on Standards of Training, and the Certification and Watchkeeping for Seafarers, and the newly implemented International Code for Ships Operating in Polar Waters, or Polar Code (IMO, 2017).

The Polar Code of 2017 sets new standards for vessels travelling in polar areas to mitigate environmental damage and improve safety (IMO, 2017). The Polar Code, however, currently excludes fishing vessels and vessels on government service, thereby excluding many shipping activities, particularly in the Antarctic region (IMO, 2017). Many ships travelling these waters will therefore continue to pose risks to the environment and to themselves, as they are not regulated under the Polar Code *(high confidence).* The Polar Code does not enhance enforcement capabilities or include environmental protection provisions to address a number of particular polar region-specific risks such as black carbon, ballast water, and heavy fuel oil transport and use in the Arctic (Anderson, 2012; Sakhuja, 2014; IMO, 2017). However, both Russian and Canadian legislation provide the possibility for stricter shipping provisions in ice-covered waters. The IMO has prohibited the use of heavy fuel oil in the Antarctic.

States can individually or cooperatively pursue the establishment of Special Areas and Particularly Sensitive Sea Areas at the IMO with a view to protect ecologically-unique or -vulnerable and economically- or culturally-important areas in national and international waters from risks and impacts of shipping, including through routing, discharge and equipment measures. Continued, and in some areas, greater international cooperation on shipping governance can facilitate addressing emerging climate change issues (Arctic Council, 2015a; ARR, 2016; PEW Charitable Trust, 2016; Chénier et al., 2017; IMO, 2017) (*high confidence*). Cooperation of the member states of the Arctic Council resulted in the 2011 Agreement on Cooperation on Aeronautical and Maritime Search and Rescue in the Arctic and in the 2013 Agreement on Cooperation on Marine Oil Pollution Preparedness and Response in the Arctic. These agreements can, if adequately implemented, reduce risks from increased Arctic shipping (*medium confidence*), however, developing more effective measures is needed as preparedness and response gaps still exist, for example, for the central Arctic Ocean.

Industry has responded to the increase in shipping activity by investing in development of shipping designs for travel in mixed-ice environments (Stephenson et al., 2011; Stephenson et al., 2013). These increases in investments 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 has 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; Lewis and Prince, 2018) (*medium confidence*). A comparative risk assessment for spills has been developed for the Arctic waters (Robinson et al., 2017) and Statoil has developed and uses risk assessment decision-support tools for environmental management, together with environmental monitoring (Utvik and Jahre-Nilsen, 2016). These tools facilitate the assessment of Arctic oil-spill response capability, ice detection in low visibility, improved management of sea ice and icebergs, and numerical modelling of icing and snow as risk mitigation.

3.5.2.8 Arctic Human Health and Well Being

At present health adaptation to climate change is generally under-represented in policies, planning, and programming (AHDR, 2014). 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 Arctic, 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 publicly available documentation on adaptation initiatives is skewed in the Arctic, with more than three-quarters coming from Canada and USA (Ford et al., 2014a; Loboda, 2014).

Many Arctic 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 national adaptation strategy outlines 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 a 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 the Arctic (Ford et al., 2014a; Ford et al., 2014b). Adaptation at the local scale is broad, ranging 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; Brubaker et al., 2011; Harper et al., 2012; Brubaker et al., 2013; Douglas et al., 2014; Austin et al., 2015; Bunce et al., 2016; Cunsolo 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). A new initiative to operationalise One Health concepts and approaches under the AC's Sustainable Development Working Group has gained momentum since 2015 (Ruscio et al., 2015). One Health approaches seek to link human, animal, and environmental health, using interdisciplinary and participatory methods that can draw on indigenous knowledge and local knowledge (Dudley et al., 2015). Thus far, the initiative has supported new regional-to-international networks, and proposals for its expansion. In the future, the ability to manage, respond, and adapt to climate-related health challenges will be a defining issue for the health sector in the Arctic (Ford et al., 2010; Durkalec et al., 2015) (*medium confidence*).

Table 3.4: Response of key human sectors /systems to climate change in polar regions. Table 3.4 summarizes the consequences, interacting drivers, responses, and assets of climate change responses by select human sectors (i.e., social-ecological systems) 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 Paris 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.

Sector /System	*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 eastern Bering Sea and Barents Sea as well as the CAMLR Convention area.	Changes in human preference, demand, and markets, changes in gear, changes in policies affecting property rights. Changes due to offshore development and transportation.
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 seriously 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 ensure free-range movements;	Flexibility in movement to respond to changes in pastures, secure land use rights and adaptive management. Continued	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.

		supplemental feeding.	economic viability and cultural tradition.		
Tourism (Arctic and Antarctic)	Warmer conditions, more open water, Public perception of 'last chance' opportunities	Increased visitation, (quantity and quality) increase in off-season tourism to polar regions	Policies to ensure 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
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 production levels, ice roads, flooding events, and infrastructure	Some shifts in practices, greater interest in offshore 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)
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. Shortening windows of operation for use of ice roads; construction of all-season roads.	Weak regional and national economies, other disasters that divert resources, disinterest by southern-based law makers
Marine 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.	Political conflict in other areas that impeded acceptance of policies for safety requirements, timing, and movements. Changing insurance premiums.

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
Coastal settlements (See Cross-Chapter Box 9)	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

3.5.3 Governance

3.5.3.1 Local-to National Governance

Responses to climate change at and across local, regional, and national levels occur 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.

A comprehensive literature review of 157 discrete cases of Arctic adaptation initiatives Ford et al. (2014b) found that adaptation is primarily local and motivated by reducing risks and their related vulnerabilities (*high confidence*). 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 should be made sensitive to community perspectives, information should be packaged and communicated in ways that are accessible to non-experts, and processes of engagement that foster creative problem solving be used. Furthermore, success of local government involvement in adaptation planning has been closely linked to transnational municipal networks that foster social learning and in which local governments assume a role as key players (Sheppard et al., 2011; Fünfgeld, 2015) (*medium confidence*). While transnational networks can be a catalyst for action and promoting innovation, there remain outstanding challenges in measuring the effectiveness of these networks (Fünfgeld, 2015).

Adaptation through formal institutions by Indigenous people is enabled through self-government, land claims, and co-management institutions (Baird et al., 2016; Huet et al., 2017). However, organizational capacity is often a limiting factor in involvement (AHDR, 2014; Ford et al., 2014b; Forbes et al., 2015) (*high confidence*). Interactions across scales are 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 local knowledge and indigenous knowledge in governance (Cross-Chapter Box 4 in Chapter 1) (AHDR, 2014; Ford et al., 2014b; Forbes et al., 2015) (*high confidence*).

At a more regional level, Alaska's 'Climate Action for Alaska' was reconstituted in 2017 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. The role of cross-scale boundary organizations 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 has proven important in Norway (Dannevig and Aall, 2015).

At the national level, Norway, Sweden, and Finland have engaged in the European Climate Adaptation Platform ('Climate-ADAPT'), a partnership that aims to support Europe in adapting to climate change by helping 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. The Canadian government's actions on climate change have been among the most extensive of the Arctic nations, including funding of ArcticNet, a Network of Centres of Excellence, and consideration of climate change by The Northern Contaminants and Nutrition North Canada programs.

3.5.3.2 International Climate Governance and Law: Implications for International Cooperation

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) (*medium confidence*).

In both polar regions, cooperative approaches to regional governance have been developed to allow for the participation of non-state actors. In some cases, regimes allow for a substantial level of participation by specific groups of the civil society, such as stakeholders (Jabour, 2017; Keil and Knecht, 2017). For example, in the Antarctic Treaty System, the Antarctic Treaty Parties included the Scientific Committee on Antarctic Research into their Protocol on Environmental Protection to the Antarctic Treaty. 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 has contributed to modifying the balance between the interests of state and non-state 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 is inadequate when applying a precautionary approach at the regional level (*medium confidence*). For example, several studies have shown that the Convention on the Protection of the Marine Environment of the North East Atlantic (OSPAR), which applies only to the North East Atlantic, and that provides a framework for implementation of the United Nations Convention on the Law of the Sea (UNCLOS) and the Convention on Biological Diversity (CBD), are insufficient to deal with risks when 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 hydrocarbons. With ice retreating and thinning, and improved access to natural resources, coastal states are increasingly recurring to the option to invoke Article 76 of the UNCLOS (Art. 76 UNCLOS; Verschuuren, 2013) and seek to demonstrate with scientific data, submitted to the Commission on the Limits of Continental Shelf, and within a set timeline, that their continental shelf is extended. In that case they can enjoy sovereign rights beyond the Exclusive Economic Zone. It is *very unlikely* that this new trend from states to refer to Article 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, cooperation in general does occur via UNCLOS, the Convention for the Safety of Life at Sea and the Convention for the Prevention of Pollution from Ships and the Polar Code. Global environmental and climate regimes that are implemented and managed through regional regimes (such as the Kyoto Protocol or the Paris Agreement) are also relevant for the Antarctic Treaty and its Protocol on Environmental Protection, which requires, amongst other issues, a minimization of adverse environmental impacts. Cooperation in the Antarctic also occurs through the Convention on the Conservation of Antarctic Marine Living Resources (CCAMLR). Climate change and its consequences for the marine environment are a central issue for CCAMLR because it challenges ways to regulate and manage fisheries and designate and manage Marine Protected Areas. Nevertheless, CCAMLR has not agreed to any climate change program and at its most recent meeting, there was again no agreement to do so (Brooks et al. (2018), CCAMLR Report on the Thirty-seven Meeting of the Commission, CCAMLR (2018)).

3.5.3.2.1 Formal arrangements: polar conventions and institutions

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., 2016a; Wehrmann, 2016; Young, 2016). The Council is composed of eight Arctic States and six Permanent Participants representing organisations of Arctic Indigenous peoples. Observers status is open to: non-Arctic states, intergovernmental and inter-parliamentary organizations, global and regional non-governmental organizations (NGOs). The Council is an example of cooperation through soft law, a unique institutional body that does not possess a legal personality and is neither an international law nor a completely state-centric institution. However, it is acting state-centric and 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, China, South Korea, Italy, Japan, India and Singapore joined France, Germany, the Netherlands, Poland, Spain and the UK as Observers states to the Arctic Council; Switzerland was granted Observer status in 2017. Non-Arctic States are stimulating the Council towards adopting a new approach for Arctic governance that would leave greater space for their participation.

Despite lacking the role to enact hard law, three binding agreements were negotiated under the auspices of the Council (in its task forces), the latest of which is the Agreement on Enhancing International Arctic Scientific Cooperation, which is an indication that the Council is preparing a regulatory role to respond 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, 2016), 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 requires 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 Ecosystem Approach, it does not have a mandate to manage climate related risks and impacts, or apply a precautionary approach, on fisheries issues.

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 it goes through restructuring and reconfiguration (Stokke, 2013; Baker and Yeager, 2015; Pincus and Speth, 2015; Cassotta et al., 2016; Tesar et al., 2016a; 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 development for a new agreement on the Conservation and Sustainable use of Marine Biodiversity of Areas beyond National Jurisdictions (BBNJ) under the UNCLOS (Baker and Yeager, 2015; De Lucia, 2017; Nengye et al., 2017; Koivurova and Caddell, 2018) (*medium confidence*).

The Antarctic Treaty System

The Antarctic Treaty System (ATS) is the collective term for the Antarctic Treaty and related agreements. The ATS regulates international relations with respect to Antarctica. 54 countries have acceded to the Treaty and 29 of them participate in decision making as Consultative Parties. 27 countries are Party to the Convention for the Conservation of Antarctic Marine Living Resources (CCAMLR), and 40 have ratified the Protocol on Environmental Protection to the Antarctic Treaty (CEP). 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). By contrast, consensus is currently limiting work programme-level responses to climate change by CCAMLR (2017a), while opportunities exist to incorporate climate concerns into mechanisms for implementation and monitoring aimed to conserve ecosystems and the environment (Brooks et al., 2018).

3.5.3.2.2 Informal arrangements

The Antarctic Treaty Consultative Parties, through the Committee for Environmental Protection (CEP) and its 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. 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 nationally funded bilateral and multi-lateral projects are underway.

3.5.3.2.3 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 different 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 (CCU) 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).

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 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 CCAMLR, the Scientific Committee's 2009 report on Antarctic Climate Change and the Environment (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 to develop a Climate Change Response Work Programme (ATCM, 2017).

3.5.4 Towards Resilient Pathways

This section presents the status of practices, tools and strategies currently employed in the Arctic and or Antarctica that can potentially contribute to climate resilient pathways. Seven general strategies for building resilience have been recognized: i) maintain diversity and redundancy, ii) manage connectivity, iii) manage slow variables and feedbacks, iv) foster an understanding of social-ecological systems as complex adaptive systems, v) encourage learning and experimentation, vi) broaden participation, and vii) promote polycentric governance systems (Biggs et al., 2012; Quinlan et al., 2016) (Cross-Chapter Box 2 in Chapter 1).

The practices listed below are not inclusive of the many resilience-building efforts underway in the polar regions. Those described are well represented in the literature and have shown sufficient utility to merit further use (ARR, 2016; AMAP, 2017a; AMAP, 2017b; AMAP, 2018) (*high confidence*). Some require more refinement while others are well developed. The following sections assess the extent to which these practices operationalize resilience-building through knowledge co-production, the linking of knowledge with decision making, and implementation of resilience-based ecosystem management, considering also their application level and key facilitating conditions; a summary is presented in Table 3.5.

Table 3.5: Summary of the assessment of practices, tools and strategies that can contribute to climate resilient pathways. Practices are shown with the potential extent of their contribution to resilience building, considering also seven general strategies (Biggs et al., 2012; Quinlan et al., 2016; Cross-Chapter Box 2 in Chapter 1). Also shown is the current level of their application in polar regions and key conditions facilitating implementation.

Type of resilience-building activity	***Practices, tool, or strategy***	***Potential extent of contribution to resilience building*** *(Large-Moderate-Limited)* *and* ***Areas of potential contributions to resilience****:* *DIV = Maintain diversity & redundancy* *CON = Manage connectivity* *PAR = Broaden participation* *LEA = Encourage learning & experimentation* *SYS = Foster complex system understanding* *GOV = Enhance polycentric governance* *SLO = Manage slow variables and feedbacks*	***Current level of application in polar regions*** *(High-Medium-Low)* *and* ***Key conditions facilitating implementation****:* *F= Financial support* *I= Institutional support* *T&S= Technical and science support* *L&I= Local & indigenous capacity and knowledge* *C= Interdisciplinary and/or cross-cultural cooperation*

		Confidence regarding potential contribution to resilience building: ▲▲▲=high ▲▲=medium ▲=low	
Knowledge Co-Production and Integration	**Community-based monitoring**	DIV, PAR, SYS ▲▲	F, I, T&S, L&I, C
	Understanding regime shifts	LEA, SYS, SLO ▲▲▲	I, T&S, C
	Indicators of resilience and adaptive capacity	PAR, LEA, SYS, SLO ▲▲	F, L&I, T&S
Linking Knowledge with Decision Making	**Participatory scenario analysis and planning**	PAR, LEA, SYS ▲▲	T&S, L&I, C
	Structured decision making	PAR, LEA, SYS ▲	I, T&S, C
Resilience-based Ecosystem Stewardship	**Adaptive ecosystem governance**	DIV, PAR, LEA, SYS, GOV, SLO ▲▲▲	I, T&S, L&I, C
	Spatial planning for biodiversity	DIV, CON, GOV, SLO ▲▲	I, T&S, L&I, C
	Linking ecosystem services with human livelihoods	DIV, PAR, SYS, GOV, SLO ▲▲▲	I, T&S, L&I, C

3.5.4.1 Knowledge Co-production and Integration

The co-production of knowledge and transdisciplinary research are currently contributing to the understanding of polar climate change through the use of a diversity of cultural, geographic, and disciplinary perspectives that provide a holistic framing of problems and possible solutions (Miller and Wyborn, 2018; Robards et al., 2018) (*high confidence*).

Several factors are important in successful knowledge co-production, including use of social-ecological frameworks, engagement of a broad set actors with diverse epistemological orientations, a 'team science' approach to studies, strong leadership, attention to process (vs only products), and mutual respect for cultural differences (Meadow et al., 2015; National Research Council, 2015; Petrov et al., 2016) (*high confidence*). Knowledge co-production involving Indigenous peoples comes with its own set of challenges (Armitage et al., 2011; Robards et al., 2018). While advancements have been made, the practice of knowledge co-production would benefit from further experimentation and innovation in methodologies and better training of researchers (van der Hel, 2016; Vlasova and Volkov, 2016; Berkes, 2017) (*medium confidence*). Three aspects of knowledge co-production are highlighted below.

3.5.4.1.1 Community-based monitoring

Community-based monitoring (CBM) in the Arctic has emerged as a practice of great interest because of its potential to link western ways of knowing with local knowledge and indigenous knowledge (Retter et al., 2004; Johnson et al., 2015a; Johnson et al., 2015b; Kouril et al., 2016; AMAP, 2017a; Williams et al., 2018). In several CBM programs, innovative approaches using the internet, mobile phones, hand-held information devices (PDAs), and camera-equipped GPS units are capturing, documenting and communicating local observations of change (Brubaker et al., 2011; Brubaker et al., 2013). The integration of community observations with instrument-based observations and its use in research has proven challenging, with technical and cultural issues (Griffith et al., 2018). Execution of CBM programs in the Arctic has also proven to be labour intensive and difficult to sustain, requiring long-term financial support, agreements specifying data ownership, sufficient human capital, and in some cases, the involvement of boundary organizations that provide technical support (Pulsifer et al., 2012; Eicken et al., 2014) and link CBM with governance (CAFF, 2015b; Robards et al., 2018). As is the case in all knowledge production, power relationships (i.e., who decides what is a legitimate observation, who has access to resources for involvement, who benefits) have been challenging where the legitimacy of local knowledge and indigenous knowledge is questioned (e.g., Pristupa et al., 2018). There is *high agreement* and *limited evidence* that CBM facilitates knowledge co-production and resilience building. More analyses of Arctic communities and their institutional capabilities related to CBM are needed to evaluate the potential of these observation systems, and experimentation and innovation may help determine how CBM can more effectively inform decision making beyond the community (Johnson et al., 2015a; Johnson et al., 2015b) (*medium confidence*).

3.5.4.1.2 Understanding regime shifts

Regime shifts are especially important in polar regions where there are limited data and where rapid directional change suggests the possibility of crossing thresholds that may dramatically alter the flow of ecosystem services (ARR, 2016). Better understanding of the thresholds and dynamics of regime shifts (i.e. SES state changes) is especially important for resilience building (ARR, 2016; Biggs et al., 2018; Rocha et al., 2018) (*high confidence*). While polar regime shifts have been documented (Biggs et al., 2018), most are poorly understood and rarely predictable (Rocha et al., 2018) (*high confidence*). Moreover, the focus on Arctic regime shifts to date has been on almost entirely on biophysical state changes that impact social systems. A limited number of studies have examined social regime shifts and fewer the feedbacks of social regimes shifts on ecosystems (Gerlach et al., 2017). Future needs for advancing knowledge of regime shifts include: 1) continued and refined updating of details on past regimes shifts, 2) structured comparative analysis of these phenomena to ascertain common patterns and variation, 3) greater investment in research resources on potential large-scale regime shifts, and 4) great attention on how social and economic change may affect ecosystems (ARR, 2016; Biggs et al., 2018).

3.5.4.1.3 Indicators of resilience and adaptive capacity

Well-crafted and effectively communicated indicators of polar geophysical, ecological and human systems have the potential to make complex issues more easily understood by society, including local residents and policy makers seeking to assess the implication of climate change (Petrov et al., 2016; Carson and Sommerkorn, 2017) (*medium confidence*). Having indicators of change is no guarantee they will be used; access to information, awareness of changing conditions, and the motivation to act are also important (e.g., van der Linden et al., 2015).

Indicators of the state of polar geophysical systems, biodiversity, ecosystems, and human well-being are monitored as part of polar programs. For example, indicators are reported by the Arctic Council working groups Arctic Monitoring and Assessment Programme and Conservation of Arctic Flora and Fauna (e.g., Odland et al., 2016; CAFF, 2017; Box et al., 2019), the International Arctic Social Science Association (e.g., AHDR, 2014), the CCAMLR Ecosystem Monitoring Programme (e.g., Reid et al., 2005) and the Southern Ocean Observing System (e.g., Meredith et al., 2013).

There is limited development of indicators of social-ecological resilience (Jarvis et al., 2013; Carson and Sommerkorn, 2017). As well, indicators of human adaptive capacity are typically based on qualitative case studies with limited quantitative data, and thus have limited comparability and generalizability (Ford and King, 2013; Petrov et al., 2016; Berman et al., 2017) (*high confidence*). The identification and on-going use of indicators of social-ecological resilience are theoretically best achieved through highly participatory processes that engage stakeholders of a locale, with those processes potentially resulting in self-reflection and actions that improve adaptive capacity (Quinlan et al., 2016; Carson and Sommerkorn, 2017), however, this is untested empirically (*low confidence*).

3.5.4.2 Linking Knowledge with Decision Making

While there is a growing expectation in polar (and other) regions for a more deliberate strategy to link science with social learning and policy making about climate change (and other matters) through iterative interactions of researchers, managers, and other stakeholders, meeting that expectation is confounded by several deeply rooted issues (Armitage et al., 2011; ARR, 2016; Tesar et al., 2016b; Baztan et al., 2017; Forbis Jr and Hayhoe, 2018) (*medium confidence*).

In spite of the development of practices like those described above and the establishment of many co-managed arrangements in polar regions, scientists and policy makers often work in separate spheres of influence, tend to maintain different values, interests, concerns, responsibilities and perspectives, and gain limited exposure to the other's knowledge system (see Liu et al., 2008; Armitage et al., 2011). Information exchange flows unequally, as officials struggle with information overload and proliferating institutional voices, and where local residents are mistrusting of scientists (Powledge, 2012). Inherent tensions between science-based assessment and interest-based policy, and many existing institutions often prevent direct connectivity. Further, the longstanding science mandate to remain 'policy neutral' typically leads to norms of constrained interaction (Neff, 2009) (*high confidence*).

Creating pathways towards greater climate resilience will, therefore, depend, in part, on a redefined 'actionable science' that creates bridges supporting better decisions through more rigorous, accessible, and engaging products, while shaping a narrative that instils public confidence (Beier et al., 2015; Fleming and Pyenson, 2017) (*high confidence*). Stakeholders of polar regions are increasingly using a suite of creative tools and practices for moving from theory to practice in resilience building by informing decision making and fostering long-term planning (Baztan et al., 2017). As noted above, these practices include participatory scenario planning, forecasting for stakeholders, and use structured decision making, solution visualization tools, and decision theatres (e.g., Schartmüller et al., 2015; Kofinas et al., 2016; Garrett et al., 2017; Holst-Andersen et al., 2017; Camus and Smit, 2018). The extent to which these practices can contribute to resilience building in the future will depend, in part, on the willingness of key actors, such as scientists, to provide active decision-support services, more often than mere decision-support products (Beier et al., 2015). While progress has been made in linking science with policy, more 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 will be important (*high confidence*).

3.5.4.2.1 Participatory scenario analysis and planning

Participatory scenario analysis is a quickly-evolving and widely-used practice in polar regions, and has proven particularly useful for supporting climate adaptation at multiple scales when it uses a social-ecological perspective (ARR, 2016; AMAP, 2017a; Crépin et al., 2017; Planque et al., 2019) (*medium confidence).* While there are technical dimensions in scenario analysis and planning (e.g., the building of useful simulation models that capture and communicate nuanced social-ecological system dynamics such as long-fuse big bang processes, pathological dynamics, critical thresholds, and unforeseen processes (Crépin et al., 2017), there are also creative aspects, such as the use of art to help in the visualization of possible future (e.g., Planque et al., 2019).

Participatory scenario analysis has been applied to various problem areas related to climate change responses in the polar regions. Applications demonstrate the utility of the practice for identifying possible local futures that consider climate change or socio-economic pathways (e.g., in Alaska, Ernst and van Riemsdijk, 2013; and in Eurasian reindeer-herding systems, van Oort et al., 2015; Nilsson et al., 2017) and interacting drivers of change (e.g., in Antarctica; Liggett et al., 2017). Scenario analysis proved helpful for stakeholders with different expertise and perspectives to jointly develop scenarios to inform ecosystem-based management strategies and adaptation options (e.g., in the Barents region; Nilsson et al., 2017; Planque et al., 2019) and to identify research needs (e.g., in Alaska; Vargas-Moreno et al., 2016), including informing and applying climate downscaling efforts (e.g., in Alaska; Ernst and van Riemsdijk, 2013).

A review of scenario analysis in the Arctic, however, found that while the practice is widespread and many are using best-practice methods, less than half scenarios programs incorporated climate projections and that those utilizing a backcasting approach had higher local participation than those only using forecasting (Flynn et al., 2018). It noted that integrating different knowledge systems and attention to cultural factors influence program utility and acceptance. Planque et al. (2019) also found that most participating stakeholders had limited experience using scenario analysis, suggesting the importance of process methods for engaging stakeholders when exploring possible, likely, and desirable futures. The long-term utility of this practice in helping stakeholders engage with each other to envision possible futures and be forward-thinking in decision making will depend on the science of climate projections, further development of decision support systems to inform decision makers, attention to cultural factors and worldview, as well as refinement of processes that facilitate participants' dialogue (*medium confidence*).

3.5.4.2.2 Structured decision making

Structured decision making (SDM) is an emerging practice used with stakeholders to identify alternative actions, evaluate trade-offs, and inform decisions in complex situations (Gregory et al., 2012). Few SDM processes have been undertaken in polar regions, with most as exploratory demonstration projects led by researchers. These have included indigenous residents and researchers identifying trade-offs and actions related to subsistence harvesting in a changing environment (Christie et al., 2018) stakeholder interviews to show how a 'triage method' can link community monitoring with community needs and wildlife management priorities (Wheeler et al., 2018), and the application of multicriteria decision analysis to address difficult decisions related to mining opportunities in Greenland (Trump et al., 2018). The Decision Theater North at the University of Alaska is also being explored as an innovative method of decision support (Kofinas et al., 2016). SDM may have potential in creating climate resilience pathways in polar regions (*low confidence*), but there is currently limited experience with its application.

3.5.4.3 Resilience-based Ecosystem Stewardship

Renewable resource management and biodiversity conservation that seek to maintain resources in historic levels and reduce uncertainty before taking action remains the dominant paradigm in polar regions (Chapin III et al., 2009; Forbes et al., 2015). The effectiveness of this approach, however, is increasingly challenged as the ranges and populations of species and state of ecosystems are being affected by climate change (Chapin III et al., 2010; Chapin III et al., 2015). Three practices that build and maintain social-ecological resilience in the face of climate change include Adaptive Ecosystem Governance, Spatial Planning for Biodiversity, and Linking Management of Ecosystem Services with Human Livelihoods.

3.5.4.3.1 Adaptive ecosystem governance

'Adaptive Ecosystem Governance' differs from conventional resource management or integrated ecosystem management (Chapin III et al., 2009; Chapin III et al., 2010; Chapin III et al., 2015), with a strong 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). This approach strengthens response options by maintaining or increasing resource diversity (to support human adaptation) and biological diversity (to support ecosystem adaptation) (Biggs et al., 2012; Chapin III et al., 2015; Quinlan et al., 2016) (*high confidence*). Adaptive ecosystem governance emphasizes iterative social learning processes of observing, understanding, and acting with collaborative partnerships, such as adaptive co-management arrangements currently used in regions of the Arctic (Armitage et al., 2009; Dale and Armitage, 2011; Chapin III et al., 2015; Arp et al., 2019). This approach is also currently realized through adaptive management of Arctic fisheries in Alaska that combines annual measures and within-season provisions informed by assessments of future ecosystem trends (Section 3.5.2.1), and the use of simulation models with Canadian caribou co-management boards to assess the cumulative effects of proposed land-use change with climate change (Gunn et al., 2011; Russell, 2014a; Russell, 2014b). Linking these regional efforts to pan-polar programs can enhance resilience building cross multiple scales (e.g., Gunn et al., 2013) (*medium confidence*).

3.5.4.3.2 Spatial planning for biodiversity

Shifts in the distribution, abundance and human use of species and populations due to climate-induced cryosphere and ocean change, concurrent with land-use changes, increase the risks to ecosystem health and biodiversity (Kaiser et al., 2015). Building resilience in these challenging conditions follows from spatial planning for biodiversity that links multiple scales and considers how impacts to ecosystems may materialize in social-ecological systems elsewhere (Bengtsson et al., 2003; Cumming, 2011; Allen et al., 2016). Developing pathways for spatial resilience in polar regions involves systematic planning and designating networks of protected areas to protect connected tracts of representative habitats, and biologically and ecologically significant features (Ban et al., 2014). Protected area networks that combine both spatially rigid and spatially flexible regimes with climate refugia can support ecological resilience to climate change by maintaining connectivity of populations, foodwebs, and the flow of genes across scales (McLeod et al., 2009). This approach reduces direct pressures on biodiversity, and thus, gives biological communities, populations, and ecosystems the space to adapt (Nyström and Folke, 2001; Hope et al., 2013; Thomas and Gillingham, 2015) (*medium confidence*). Networks of protected areas are now being planned (Solovyev et al., 2017) and implemented (Juvonen and Kuhmonen, 2013) in the marine and terrestrial Arctic, respectively; expanding the terrestrial protected area network in Antarctica is discussed (Coetzee et al., 2017). The planning of protected area networks in polar regions is currently an active topic of international collaboration in both polar regions (Arctic Council, 2015b; CCAMLR, 2016a; Wenzel et al., 2016). Designating marine protected area networks contributes to achieving Sustainable Development Goal 14 and the Aichi Targets of the Convention for Biological Diversity but is often contested due to competing interests for marine resources.

3.5.4.3.3 Linking eosystem services with human livelihoods

Incorporating measures of ecosystem services into assessments is key in integrating environmental, economic, and social policies that build resilience to climate change in polar regions (CAFF, 2015a; Malinauskaite et al., 2019; Sarkki and Acosta García, 2019) (*high confidence*). Currently, there is limited recognition of the wide range of benefits people receive from polar ecosystems and a lack of management tools that demonstrate their benefits in decision-making processes (CAFF, 2015a). The concept of ecosystem services is increasingly used in the Arctic, yet there continues to be significant knowledge gaps in mapping, valuation, and the study of the social implications of changes in ecosystem services. There are few Arctic examples of the application of ecosystem services in management (Malinauskaite et al., 2019). A strategy of ecosystem stewardship, therefore, is to maintain a continued flow of ecosystem services, recognizing how their benefits provide incentives for preserving biodiversity, while also ensuring options for sustainable development and ecosystem-based adaptation (Chapin III et al., 2015; Guerry et al., 2015; Díaz et al., 2019). Arctic stewardship opportunities at landscape, seascape, and community scales to a great extent lie in supporting culturally engrained (often traditional indigenous) values of respect for land and animals, and reliance on the local environment through the sharing of knowledge and power between local users of renewable resources and agencies responsible for managing resources (Mengerink et al., 2017) (*high confidence*). In the Antarctic, ecosystem stewardship is dependent on international formally-defined and informally-enacted cooperation, and the recognition of its service to the global community (Section 3.5.3.2).

3.6 Synopsis

This chapter has assessed the consequences of climate change in the polar regions in three sections, focusing on sea ice and the ocean (Section 3.2), glaciers and ice sheets (Section 3.3), and permafrost and snow on land (Section 3.4). A systems approach was taken to assess individual and interacting changes within and between these elements to consider consequences, impacts and risks for marine and terrestrial ecosystems and for people. Mapping on to those observed and projected impacts, Section 3.5 assessed human responses to climate change in the polar regions. This brief synopsis considers the chapter findings across sections, and draws out three key assessment points that inform global responses to polar ocean and cryosphere change.

1) Climate-induced changes to the polar cryosphere and oceans have global consequences and impacts.

Loss of Arctic sea ice continues to add to global radiative forcing and has the potential to influence midlatitude weather on timescales of weeks to months (Section 3.2.1 Box 3.2), the Southern Ocean takes up a disproportionately high amount of atmospheric heat and carbon (Section 3.2.1), melting polar glaciers and the Antarctic and Greenland ice sheets contribute to observed and projected sea level rise long into the future (Sections 3.3.1, 3.3.2), and projected widespread disappearance of permafrost has the potential to accelerate global warming through the release of carbon dioxide and methane (Sections 3.4.2, 3.4.3).

2) Across many aspects, the polar regions of the future will appear significantly different from those of today

By the end of this century, Arctic sea ice and snow on land are projected to be diminished compared with today, as are the masses of the Antarctic and Greenland Ice Sheets and the polar glaciers (Sections 3.2.2, 3.3.2; 3.4.2). Acidification of both polar oceans will have progressed; this, and changing marine habitats associated with ocean warming, are projected to impact marine ecosystems (Sections 3.2.2, 3.2.3). Permafrost thaw and decrease in snow on land are projected to drive habitat and biome shifts, affecting ranges and abundance of ecologically-important species, and driving changes in wildfire, vegetation and human infrastructure (Sections 3.4.2, 3.4.3). Collectively, these very different future polar environments pose strong challenges to sustainable use of natural resources, infrastructure, cultures, and lives and livelihoods. Although manifested locally, these very different future polar environments have the potential to continue/accelerate the global impacts noted above.

3) Choices are available that will influence the nature and magnitude of changes, potentially limiting their regional and global impacts and increasing the effectiveness of adaptation actions.
Compared with high greenhouse gas emissions scenarios, projections using low emissions scenarios result in polar regions that will be significantly less altered. For example, for stabilised global warming of 1.5°C, a sea ice-free Arctic in September is projected to occur significantly less frequently than at 2°C (1% c.f. 10-35%) (Section 3.2.2). The potential for reduced but stabilised Arctic autumn and spring snow extent by mid-century for RCP2.6 contrasts with continued loss under RCP8.5, and the area with near-surface permafrost is projected to decrease less by 2100 under RCP2.6 than under RCP8.5 (Section 3.4.2). Polar glaciers are projected to lose much less mass between 2015 and 2100 under RCP2.6 compared with RCP8.5 (Cross-Chapter Box 6 in Chapter 2). Acidification of the polar oceans will progress more slowly and affect much smaller areas under RCP2.6 compared with RCP8.5 (Section 3.2.2). These differences have strong implications for natural resource management, economic sectors, and Arctic cultures (Section 3.5.2). The choices that enable these differences influence the rate and magnitude of polar change, their consequences and impacts at regional and global scales, the effectiveness of adaptation, and opportunities for climate resilient pathways (Section 3.5.4).

3.7 Key Knowledge Gaps and Uncertainties

Beyond this report, progress requires that future assessments demonstrate increased confidence in various key aspects; this can be achieved by narrowing numerous gaps in knowledge. Some of the critical ones, which are priorities for future initiatives, are outlined here.

Overturning circulation in the Southern Ocean is 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. Improved mechanistic understanding of the observed changes and trends in Antarctic sea ice is required, notably the decadal increase and very recent rapid retreat. This has consequences for climate, ecosystems and fisheries; however, lack of understanding and poor model performance translates to very limited predictive skill.

Trends in snow water equivalent over Arctic land are inadequately known, reducing confidence in 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 is inadequate knowledge concerning carbon dioxide and methane emissions from land and sub-sea permafrost.

There are clear regional gaps in knowledge of polar ecosystems and biodiversity, and insufficient population estimates/trends for many key species. Biodiversity projections are limited by key uncertainties regarding the potential for organisms to adapt to habitat change and the resilience of foodweb structures. 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.

There is a need to better understand the evolution of polar glaciers and ice sheets, and their influences on global sea level. Longer and improved quantifications of their changes are required, especially where mass losses are greatest, and (relatedly) better attribution of natural versus anthropogenic drivers. Better understanding of the sensitivity of Antarctica to marine ice sheet instability is required, and whether recent changes in West Antarctica represent the onset of irreversible change.

There are critical gaps in knowledge concerning interactions between the atmosphere and specific elements of the polar ocean and cryosphere. Detailed assessment of atmospheric processes was outside the remit of this chapter, however such gaps limit understanding of ongoing and future trajectories of the polar regions and their climate systems. Relatedly, there is a paucity of studies analysing differences in the trajectories of polar cryosphere and ocean systems between low and very low greenhouse gas emission scenarios.

There are critical needs to better understand the efficacy and limits of strategies for reducing risk and strengthening resilience for polar ecosystems and people, including the contribution of practices and tools to contribute to climate-resilient pathways. Knowledge on how to translate existing theoretical understandings of social-ecological resilience into decision making and governance is limited. There is limited understanding concerning the resources that are needed for successful adaptation responses and about the effectiveness of institutions in supporting adaptation. 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.

[START FAQ3.1 HERE]

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

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

Among the risks to 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, but climate change alters Arctic and Antarctic marine habitats, and affects the ability of polar species and ecosystems to withstand or adapt to physical changes. This has consequences for where, when, and how many fish can be captured. 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 most developed, scientists are exploring modifications to existing precautionary, ecosystem-based management approaches to increase the scope for adaptation to climate change impacts on marine ecosystems and fisheries.

New shipping routes through the Arctic offer cost savings because they are shorter than traditional passages via the Suez or Panama Canals. Ship traffic has already increased and is projected to become more feasible in the coming decades as further reductions in sea ice cover make Arctic routes more accessible. Increased 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 allows greater access to offshore petroleum resources and ports supporting resource extraction on land.

The polar regions influence the global climate through a number of processes. As spring snow and summer sea ice cover decrease, more heat is absorbed at the surface. There is growing evidence that ongoing changes in the Arctic, primarily sea ice loss, can potentially influence mid-latitude weather. 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.

Melting ice sheets and glaciers in the polar regions cause sea levels to rise, affecting coastal regions and their 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, ice loss from the Antarctic Ice Sheet has continued to accelerate, driven primarily by increased melting of the underside of floating ice shelves, which has caused glaciers to flow faster. 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 future sea level rise.

The Southern Ocean that surrounds Antarctica is the main region globally where waters at depth rise to the surface. Here, they become transformed into cold, dense waters that sink back to the deep ocean, storing significant amounts of human-produced heat and dissolved carbon for decades to centuries or longer, and helping to slow the rate of global warming in the atmosphere. Future changes in the strength of this ocean circulation can so far only be projected with limited certainty.

[END FAQ3.1 HERE]

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