Chapter 3: Polar Regions

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Executive Summary

Despite their locations at the opposing ends of the planet, the Polar Regions are interconnected parts of the Earth System that exert significant influence over the lives and livelihoods of humanity. This chapter assesses the state of knowledge concerning the different interdisciplinary elements of the Arctic and Antarctic systems, how they are affected by climate change and how they are likely to develop into the future, what their local, regional and global impacts might be, and the opportunities and challenges of varying response options. Key findings from this chapter are as follows.

How and why are the Polar Regions changing?

It is *virtually certain*[1] that Antarctica and Greenland have lost mass over the past decade; this mass loss occurred at accelerated rates for Greenland (*virtually certain*) and Antarctica (*very likely*). The large mass loss in Greenland is *very likely* caused by enhanced surface melt, runoff and glacier flow (*high confidence*[2]). The significant mass loss in Antarctica is very likely due to enhanced glacier flow (*high confidence*). For both ice sheets, there is *high confidence* that ocean-ice sheet interaction drives key ice sheet mass loss processes {3.2.1; 3.2.2; 3.2.4}.

There is increasing evidence that observed Antarctic Ice Sheet changes are irreversible on centennial timescales (*medium confidence*). This is especially pronounced in regions of West Antarctica such as the Amundsen Sea sector, where evidence of marine ice sheet instability has been observed and reproduced in simulations {3.2.3}.

Observed glacier mass loss over the last few decades is attributable to anthropogenic climate change with *high confidence*. Attributing changes in ice sheets to anthropogenic change remains challenging. In both Polar Regions, strong regional variability in atmospheric and oceanic circulation presently precludes unambiguous attribution of ice sheet changes to anthropogenic forcing {3.2.2, 3.2.4}.

Observed reductions in Arctic terrestrial snow and sea ice (*very high confidence*) are influencing the global climate system through albedo changes (*high confidence*). Anthropogenically-driven reductions in Arctic spring snow and summer sea ice cover have continued unabated since AR5 (*very high confidence*), with consequences for the global heat budget (*high confidence*). There is *high confidence* that reduced snow and ice cover can influence weather and climate outside of the Arctic, though presently there is *low confidence* concerning specific mechanisms and the extent to which this influence has been manifested {3.4.1.1; Box 3.1}.

It is *virtually certain* that projected warming will result in continued loss of Arctic sea ice in summer and terrestrial snow in spring, due to their strong sensitivity to temperature forcing. At 2°C stabilized global warming (RCP4.5), occurrence of at least one sea ice-free Arctic summer every 5 years is *very likely*; for stabilized global warming at 1.5°C, summer sea ice is *very likely* to survive albeit with greatly reduced extent. Important differences in Arctic spring snow cover projections between scenarios emerge by end of century, with stabilized snow loss for RCP4.5 but continued loss for RCP8.5 (*high confidence*). Antarctic sea ice trends have shown a weak response to greenhouse gas-driven warming, though with strong observed reductions since 2015. Known model biases and disagreement with observed trends mean there is currently *low confidence* in climate model projections of Antarctic sea ice change {3.3.1.1; 3.3.2.1}.

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[1] FOOTNOTE: In this Report, the following terms have been used to indicate the assessed likelihood of an outcome or a result: Virtually certain 99–100% probability, Very likely 90–100%, Likely 66–100%, About as likely as not 33–66%, Unlikely 0–33%, Very unlikely 0–10%, Exceptionally unlikely 0–1%. Additional terms (Extremely likely: 95–100%, More likely than not >50–100%, and Extremely unlikely 0–5%) may also be used when appropriate. Assessed likelihood is typeset in italics, e.g., *very likely* (see Section 1.8.3 and Table 1.2 for more details).

[2] FOOTNOTE: In this Report, the following summary terms are used to describe the available evidence: limited, medium, or robust; and for the degree of agreement: low, medium, or high. A level of confidence is expressed using five qualifiers: very low, low, medium, high, and very high, and typeset in italics, e.g., *medium confidence*. For a given evidence and agreement statement, different confidence levels can be assigned, but increasing levels of evidence and degrees of agreement are correlated with increasing confidence (see Section 1.8.3 and Table 1.2 for more details).

Observations of increased baseflow in northerly-flowing Arctic rivers over the last several decades are attributable to permafrost thaw and a concomitant enhancement in groundwater discharge (*high confidence*). Other Arctic freshwater systems have also changed, including decreased lake ice cover duration and regionally-variable surface wetting/drying due to intensified thermokarst activity**,** but there is only *medium confidence* in these trends because observations are not comprehensive in space and time {3.4.1.2}.

Changes in permafrost influence the global climate system through emissions of the greenhouse gases carbon dioxide and methane released from the microbial breakdown of organic carbon. The organic carbon pool stored in Arctic and boreal permafrost soils contains almost twice the amount of carbon as the atmosphere (*high confidence*). Environmental changes including observed increases in permafrost temperature cause the mobilisation of this organic carbon (*high confidence*). Models project continued substantial loss of permafrost carbon by 2100 under RCP8.5 (*medium confidence*), while emission scenarios limiting global mean temperature rise to 2 degrees (e.g. RCP4.5) will reduce carbon emissions from permafrost (*high confidence*) {3.4.1.3; 3.4.2.3; 3.4.3.1}.

Drawdown of atmospheric carbon by the Southern Ocean has become reinvigorated since the early 2000s (*high confidence*), with consequences for the global carbon budget. This follows a period spanning the 1990s during which the Southern Ocean sink for carbon exhibited a marked weakening. Factors contributing to these decadal changes include changes in the strength of the winds that overlie the Southern Ocean, differing regional changes in surface ocean temperature, and fluctuations in the rate of oceanic overturning circulation. This sink represents up to half of the global ocean uptake of carbon from the atmosphere, with strong implications for climate change and rates of ocean acidification {3.3.1.2; 3.3.1.3; 3.3.2}.

The polar oceans are changing more rapidly than the global ocean as a whole, with consequences for marine productivity (*high confidence*). The heat stored in the polar oceans has increased in recent decades, contemporaneously with pronounced changes in pH, stratification, freshwater content and sea ice cover. Arctic sea ice loss has led to an increase in primary production, spring plankton blooms being dominated by larger-celled organisms, and increased incidence of fall blooms. In the Southern Ocean, model projections indicate increases in primary production up to 2100 under RCP8.5, as a result of cryospheric changes and increased temperature {3.3.1; 3.3.3}.

The distributions of some ecologically- and commercially-important species in both Arctic and Southern Ocean ecosystems have changed in response to climate-driven changes in the ocean and sea ice (*high confidence*). This includes habitat expansion of key boreal fish stocks in the Barents Sea in the European Arctic (*high confidence*). Climate projections indicate further range shifts in the future (*medium confidence*)**,** including to Antarctic krill, a cornerstone species in Southern Ocean foodwebs that is commercially exploited {3.3.3}.

Why do the changes in the Polar Regions matter, regionally and globally, and to whom?

Observed changes in ice sheets and glaciers raise sea level worldwide. Since AR5, it has become *virtually certain* that melting ice sheets and glaciers dominate observed sea level rise, and that this contribution is accelerating (*high confidence*), with consequences for human populations worldwide, especially those that live along coasts and on low-lying islands. {3.2.1}

Observed changes in Arctic snow, permafrost, and ice have consequences for northern rural and urban communities (*high confidence*). Food insecurity is increasing for Arctic peoples because of environmental changes to animal habitat and movement, changes to travel conditions to access hunting grounds, and observed declines in species important for food and local cultures (*high confidence*). Under RCP4.5, 70% of Arctic circumpolar infrastructure is located in areas where permafrost is projected to thaw by 2050 (*high confidence*). There is *very high confidence* that Arctic ship traffic has increased over the past decade, and will continue to increase in the future, which presents socio-economic opportunities, but also environmental and cultural risks for northern communities. {3.4.3.2}

Future climatic change has the potential to impact productivity of several commercially-important species in the polar oceans (*high confidence*). Both Arctic fish stocks and Southern Ocean species (including Antarctic krill) that are of great significance economically and for regional and global food security are potentially affected. Specific impacts will depend on the degree of global warming and on the responses of fisheries management. Regional differences in the marine manifestation of climate change and complex ecosystem interactions currently limit our ability to construct robust, regional, species dependent projections. {3.3.4, 3.3.3}.

Most sectors operating in the Polar Regions are experiencing the effects of climate change, with some negatively impacted in significant ways and some being able to take advantage of new opportunities (*high confidence*). Climate-induced stressors on polar oceans and the cryosphere generate cascading impacts on diverse societal aspects such as health, food and water security, commercial, subsistence and recreational fishing, reindeer herding, infrastructure, and culture. New opportunities for marine transport and tourism have emerged, but are creating additional risks for the polar environment and cultures, as development of region-specific regulatory systems is lagging behind {3.3.4, 3.4.3}.

What are options for responding to these changes, and what are the constraints on responding effectively?

Human responses to climate change and concurrent drivers of change in the Polar Regions are feedbacks that affect system trajectories and potentially mitigate or confound adaptation strategies and resilient pathways (*medium confidence*). The effects of climate change interact with land-use change, renewable and non-renewable resource industries, changing marine use, and sustainable development, which have the potential to constrain human choice, increase risk, and limit the ability to adjust behaviour {3.5.3, 3.5.4, 3.5.6}.

International cooperation responding to climate change in Polar Regions is now occurring in a new multi-level governance landscape that has strengthened but is currently fragmented (*high confidence*). New trans-national ocean climatic governance challenges and new polar interests from outside the regions are driving stronger coordination and integration among different levels and sectors of governance. Both formal and informal actors are increasingly driven to operate in this new "global" polar landscape as both rule-makers and rule-takers. Polar informal actors are increasingly playing an active role in shaping regulations {3.5.5}.

Innovative systems of governance in the Arctic allow for uniquely cooperative responses to climate change that facilitate knowledge co-production, social learning, adaptation, and building resilient pathways (*high confidence*). These institutions recognise the role of stakeholders and indigenous peoples and depend on mutual respect and trust. They provide a basis for communication of information and perspectives among parties in decision making and offer new experimentation and learning forums to explore adaptation, climate resilience and sustainable development. Knowledge co-production (e.g. interdisciplinary and indigenous and local knowledge) provides for more holistic understanding of problems and for greater perceived legitimacy of new knowledge {3.5.6}.

Synthesis

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

3.1 Introduction: Polar Regions, People and the Planet

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

Since the IPCC AR5, there has been a growing body of scientific literature, assessments and overviews pertaining to the Polar Regions. These have afforded improved understanding of the dynamics and functioning of the Polar Regions in the context of climate change, and offer new knowledge bases that have the potential to help societies as they seek to identify responses to ongoing and future changes in the ocean and cryosphere.

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

Assessments of the Polar Regions often start with their delineation; here we begin by emphasizing that the Polar Regions are two integrated parts of the Earth System. They interact with the rest of the world through ocean, atmosphere, ecological and social systems, and play key roles as important components of the global climate system (Figure 3.1).

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

Consideration of the totality of peer-reviewed scientific knowledge is a hallmark of the IPCC assessment process. Lately, there is increasing awareness of the value of considering in parallel local and indigenous knowledge in integrated assessments of climate change, specifically because the 'multiple ways of knowing' not only broaden and strengthen the evidence base but also facilitate better understanding of the challenges facing indigenous peoples, and identification and acceptance of adaptation strategies in communities across the region (see Cross Chapter Box 3). By systematically incorporating published local and indigenous knowledge in parallel with scientific knowledge, this chapter seeks to demonstrate the benefits of incorporating the multiple ways of knowing in order to better address the key issues of climatic change and its impact on the polar regions, the planet and its people.

There is great complexity within the interdisciplinary understanding of the Polar Regions, with multiple and often interacting drivers and feedbacks causing diverse, multi-faceted responses that influence physical, biological and human systems. These are outlined in detail throughout the course of this chapter; to help navigation, a schematic is provided here including pointers to chapter sections (Figure 3.1).

Reflecting the global connectivity of the Polar Regions, we purposefully adopt a flexible approach when describing their spatial footprint in relation to particular subjects or scientific disciplines. Our broad conception is that the southern Polar Region encompasses the flow of the Antarctic Circumpolar Current at least as far north as the Subantarctic Front and fully enclosing the CCAMLR Statistical Areas (CCAMLR, 2017c), whilst the marine Arctic comprises the areas of the Arctic Large Marine Ecosystems (PAME, 2013).

The terrestrial Arctic includes the areas of the northern continuous and discontinuous permafrost zone, the Arctic biome, and the parts of the boreal biome that are characterised by cryosphere elements, such as permafrost and persistent seasonal snow cover. The spatial footprints of these Polar Regions include a vast share of the world's ocean and cryosphere: they encompass surface areas equalling 20% of the of the global ocean and more than 90% of the world's continuous and discontinuous permafrost area, both of the world's ice sheets, 69% of the world's glacier area, almost all of the world's sea ice, and land areas that are entirely snow-covered during winter.

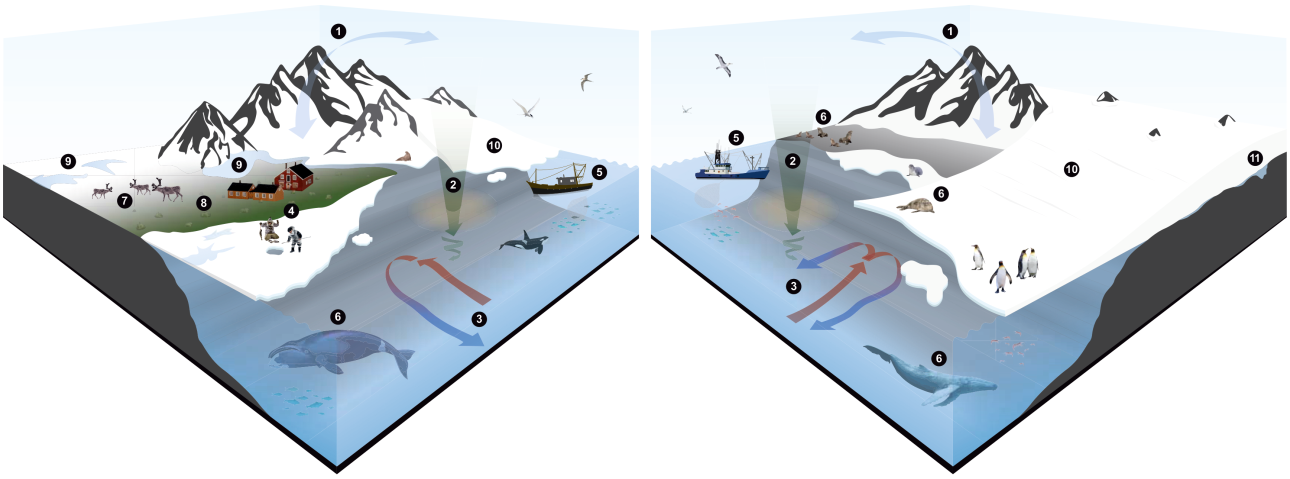


Figure 3.1: Schematic of some of the key features and mechanisms assessed in this Chapter, and by which the cryosphere and ocean in the Polar Regions influence climate, ecological and social systems in the regions and across the globe. The relevant sections wherein information can be found in this chapter are numbered. (1) The changing cryosphere influences albedo and atmospheric feedbacks, with global-scale consequences for climate {3.1; 3.4.3.1; 3.A.1}; (2) The polar oceans are key regions for the drawdown and storage of heat and carbon (including anthropogenic) from the atmosphere {3.3}; (3) Processes in the polar oceans exert strong influences on water mass formation, and driving/closure of the global ocean circulation {3.3}; (4) The Arctic is home to local and indigenous populations, whose daily life and rich and diverse cultural heritage is closely intertwined with the cryosphere {3.3.4; 3.4.3.3, 3.5}; (5) The polar regions are of increasing economic significance, bringing risks and opportunities and new challenges to cooperation, governance, and development {3.3.4, 3.4.3.3, 3.5}; (6) The polar oceans are key sites for marine biodiversity and ecosystems, with some species subject to globally-relevant commercial exploitation {3.3; 3.5} (7) The polar terrestrial regions feature unique biodiversity that is effected by changes in climate and the cryosphere, with impacts on people {3.4}; (8) Changing snow and frozen ground effects Arctic landscapes, with consequences for plants, wildlife, ecosystems, people, and global climate {3.4}; (9) Terrestrial freshwater systems influence hydrological and ecological processes on land and off shore, with impacts on northern populations {3.4}; (10) Meltwater discharged from the Greenland and Antarctic Ice Sheets exerts major influences on global sea level {3.2}; (11) Subglacial discharge has the capacity to influence ocean properties, marine productivity and the ecosystem {3.3}

[START BOX 3.1 HERE]

Box 3.1: How have Global Changes Affected Polar Regions and What are the Large-scale Feedbacks?

Arctic Amplification of Climate Change and Recent Events

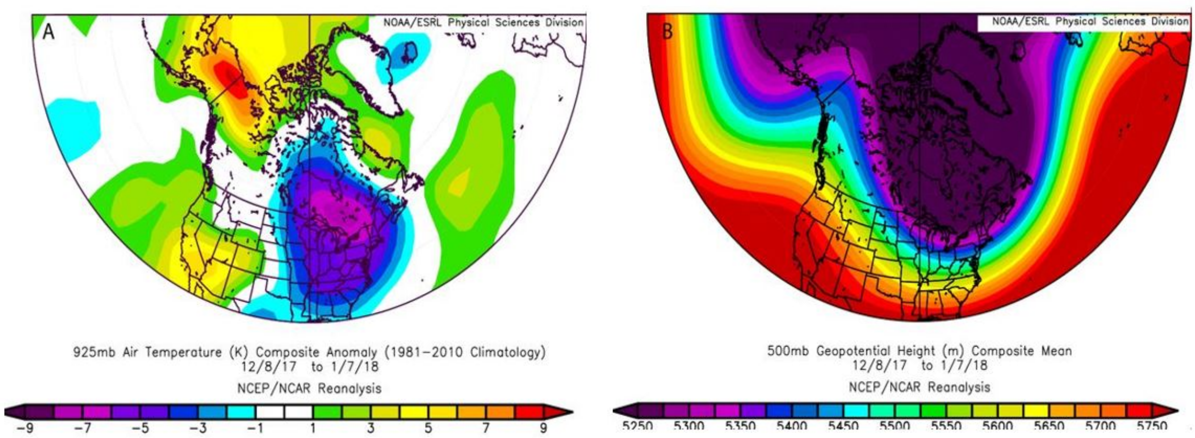
For the last two decades Arctic air temperature change are double that of the global, and a clear indicator climate change (Notz and Stroeve, 2016b; Richter-Menge et al., 2017).This ratio of 2:1 is robust for the future in at least the last two generations of global climate assessments (Kattsov and Pavlova, 2015). Mechanisms for Arctic Amplification include: reduced summer albedo due to sea ice and snow cover loss, the increase of total water vapor content in the Arctic atmosphere, a potential decrease of total cloudiness in summer and increase in winter (Makshtas et al., 2011), the additional heat generated by newly sea-ice free ocean areas that are maintained later into the autumn (Serreze and Barry, 2011), and the lower rate of heat loss to space in the Arctic relative to the sub-tropics due to lower mean temperatures (Pithan and Mauritsen, 2014).

Three recent events in the Arctic show new singular impacts. First, both winter 2016 and 2018 (Jan-Mar) had temperature anomalies in the central Arctic of +6°C, nearly double the previous record (Overland and Wang, 2016). This was caused by a split in the tropospheric vortex into two cells that advected warm air and increased moisture, separately from the Pacific and Atlantic Oceans into the central Arctic. Delayed freeze-up of sea ice in subarctic seas (Chukchi, Barents and Kara) acted as a positive feedback allowing warmer temperatures to progress further toward the North Pole. Second, not only are there low sea ice extents in summers since 2007, but now there are beginning to be low sea ice extents in sequential winters, i.e. 2016, 2017, 2018. Third, the Greenland ice sheet melted significantly earlier and faster in 2016 and 2017 than in previous years (Kintisch, 2017). Multi-year, large magnitude extreme positive Arctic temperatures and sea ice minimums since AR5, provide high agreement and medium evidence of contemporary states well outside the envelope of previous experience.

Potential for Arctic and Mid-Latitude Weather Linkages

Since AR5 understanding of Arctic and mid-latitude weather connections has become a societally important topic potentially impacting 10s of millions of people (Jung et al., 2015), but the science is difficult given the complexity of intervening meteorological processes. Assessments continue to be controversial (National Research Council, 2014; Barnes and Polvani, 2015; Francis, 2017). Arctic forcing from sea ice and snow loss and increased temperatures is clearly increasing, but the link to mid-latitude impacts is mediated by jet stream dynamics; connectivity is reduced by the influence of chaotic internal natural variability and other tropical and oceanic forcing. The potential for Arctic-mid-latitude weather linkages varies for different jet stream patterns (Grotjahn et al., 2016; Messori et al., 2016; Overland and Wang, 2018). Part of the scientific controversy is due to intra- and inter-annual intermittency in the linkage pathway.

Considerable literature exists on the potential for cold episodes in eastern Asia from Kara Sea sea-ice loss (Kim et al., 2014; Kretschmer et al., 2016). There is some analysis of cases between change in the Chukchi Sea and west of Greenland, and cold events in eastern North America (Kug et al., 2015; Ballinger et al., 2018; Overland and Wang, 2018). Such connections seem to be episodic (Cohen et al., 2018) as climatological studies do not show increases in the number of cold events in data or model projections (Ayarzaguena and Screen, 2016; Trenary et al., 2016). A potential North American example was December 2017. Warm temperatures over Alaska and record lack of sea ice (Box 3.1 Figure 1 A) helped to anchor the long wave geopotential height pattern (B), which in turn feed cold temperatures into the eastern US (A).



Box 3.1, Figure 1: A) 925 mb air temperature anomalies during December 2017. B) matching 500 mb geopotential height pattern.

Southern Hemisphere

In contrast to the Arctic, the Antarctic continent has seen less uniform temperature changes over recent decades, with warming over Western Antarctica and the Antarctic Peninsula and weak cooling over East Antarctica (e.g., Nicolas and Bromwich, 2014), though there is *low confidence* in these changes given the sparse in situ records. There is *medium confidence* through a growing body of literuatre that variability of the tropical Pacific can strongly influence these temperature changes (Turner et al., 2016; Smith and Polvani, 2017; Clem et al., 2017a) as well as Antarctic ice shelf and glaciers (Dutrieux et al., 2014; Smith and

Polvani, 2017; Paolo et al., 2018), the SH mid-latitude circulation (Schneider et al., 2015a; Raphael et al., 2016; Clem et al., 2017a) and Antarctic sea-ice extent on year-to-year (Schneider and Deser, 2017; Stuecker et al., 2017; Turner et al., 2017b) and decade-to-decade timescales (Meehl et al., 2016; Purich et al., 2016b).

The SAM, PSA and zonal-wave 3 (see supplementary material) are the dominant atmospheric drivers of Antarctic continental and sea-ice changes. Consistent with AR5, it is *likely* that Antarctic ozone depletion has been the dominant driver of the positive trend in the Southern Annular Mode during austral summer since the 1970s (Waugh et al., 2015; Schneider et al., 2015a), however new research suggests a stronger role for tropical SSTs since 2000 (Schneider et al., 2015a).

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) and find that any impacts on the jet are strongly dependent on the season and model examined. The lack of attention is primarily since Antarctic sea-ice, has shown a small but significant *increase* in extent over the satellite era, compared to the large declines in the Arctic. The region is also sparsely populated.

Attempts to link changes in the mid-latitude circulation to long-term changes in Antarctic sea-ice extent are still a matter of debate. There is *medium confidence* that there is a two-timescale response of the Southern Ocean (Ferreira et al., 2015; Seviour et al., 2016; Seviour et al., 2017) to changes in the SAM and wind stress patterns. An initial cooling which would expand Antarctic sea ice gives way to a longer-term response of the ocean circulation to the wind changes that eventually melts the sea ice. Freshwater increases due to increased glacial melt and/or rainfall changes have also been hypothesized as leading to increased sea ice (Swart and Fyfe, 2013).

As in AR5, Antarctic sea ice is projected to decline substantially over the 21st Century but with *low confidence*. Recent studies suggest that the spread in projected sea-ice responses are intimately tied to the projected changes in the SH atmospheric circulation (Holland et al., 2017; Bracegirdle et al., 2018) which will be driven by the opposing influences of ozone recovery and increasing greenhouse gases. Time will tell whether the record low values of Antarctic sea ice in austral spring 2016 with continued declines in 2017 (Turner et al., 2017b), is a harbinger of the future response of the Antarctic region to anthropogenic warming.

[END BOX 3.1 HERE]

3.2 Changes in Polar Ice Sheets and Glaciers

3.2.1 Regional Patterns of Change

3.2.1.1 East Antarctic Ice Sheet Mass Budget Estimates, Altimetry Results, GRACE, Partitioning of the Mass Budget

Since AR5, there have been advances in collecting and summarizing data on past (200–1000 years) changes in surface mass balance (SMB) in Antarctica. Significant short-term regional accumulation anomalies have been observed in Dronning Maud Land (East Antarctica) with a positive accumulation anomaly of ~350 Gt, equivalent to ~1 mm of sea level drop, between 2009 and 2011 (Boening et al., 2012; Lenaerts et al., 2013; Welker et al., 2014). On a decadal time scale, however, a study using 76 shallow firn cores from coastal and interior Dronning Maud Land (East Antarctica) found *very likely*, with *medium confidence* that there is no trend in accumulation between 1950 and 2010 (Altnau et al., 2015).

On century time scales, Thomas and others (Thomas et al., 2017) used 49 records in East Antarctica, 7 records in the Antarctic Peninsula, and 23 in West Antarctica to conclude that Antarctica has likely experienced a growth of 7 ± 1 Gt per decade during the last 200 years and 14 ± 1 Gt per decade during the last 100 years with *low confidence*. East Antarctica contributed 10% of that growth (0.8 Gt per decade), on the interior plateau, the Weddell sea coast and Dronning Maud Land. In the Peninsula, the increase began in the 1930s and accelerated in the 1990s (Thomas et al., 2015; Goodwin et al., 2016).

On longer time scales, four cores that extend back ~1000 years suggest an accumulation decrease, which is in line with the result from 67 ice cores that found no clear trend in accumulation over most of Antarctica in the last 800 years (Frezzotti et al., 2013), except for a >10% increase since the 1960s in wet coastal regions and over the highest part of East Antarctica.

Since AR5 there has been progress in estimating the Antarctic ice mass balance using the mass budget method (MBM), repeated altimetry, and measurements of the Earth's gravity field. From the MBM, East Antarctic mass balance ranges from –35 to +13 Gt yr^{-1} for the period 1979–2016 (updated from Rignot et al., 2011) , with dynamic losses concentrated in the Wilkes Land sector (Figure 3.2). From GRACE, the mass balance in 2002-2016 is +67 ± 29 Gt yr^{-1} (Velicogna et al., 2014), while Cryosat-2 indicates –3 ± 36 Gt yr^{-1} for 2010–2013 (McMillan et al., 2014). A Bayesian hierarchical modelling applied to simultaneously determine annual trends in ice dynamics, SMB anomalies, and a time-invariant solution for glacio-isostatic adjustment (GIA) using satellite altimetry, gravimetry and elastic-corrected GPS data (Zammit- Mangion et al., 2014; Martín- Español et al., 2016) indicates a growth of +56 ± 18 Gt yr^{-1} in 2003–2013 for East Antarctica dominated by SMB anomalies. The mass balance of individual East Antarctic drainage basins varies from –39 ± 36 Gt yr^{-1} (Wilkes Land) to +28 ± 8 Gt yr^{-1} (Dronning Maud Land), indicating that regions contribute differently. Glacier acceleration and thinning in Wilkes Land is reported independently from altimetry (Flament and Rémy, 2012) and MBM (Li et al., 2016).

In summary, the mass balance of East Antarctica is *not likely* different from zero, with *medium confidence*, but important glacier changes are *likely* taking place in the Wilkes Land sector, with *high confidence*.

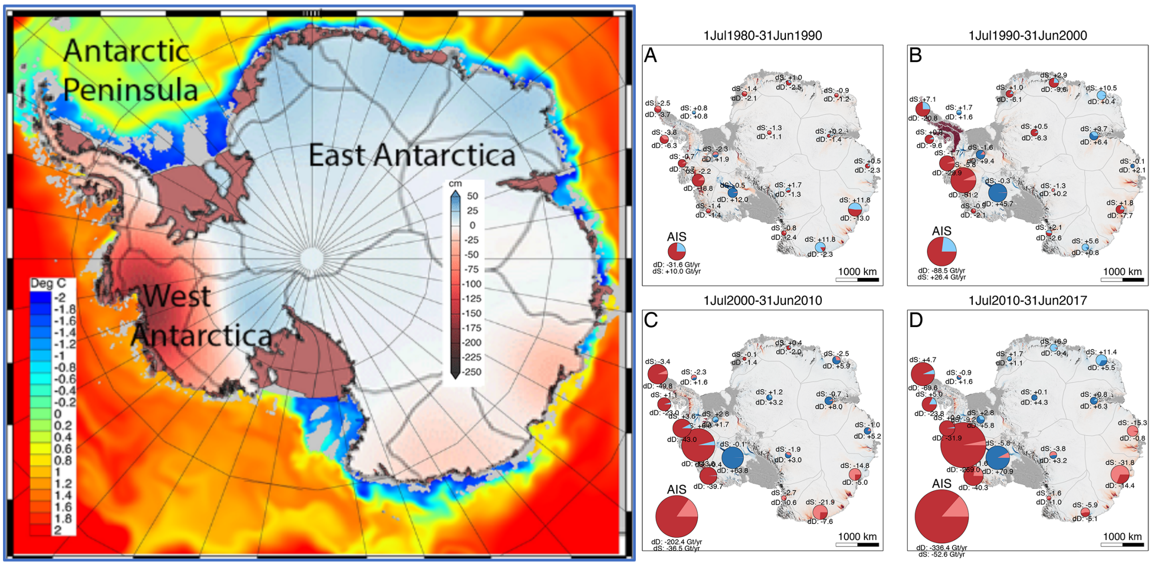


Figure 3.2: (Left) Mass balance of the Antarctic Ice Sheet for 2002–2017 from time-variable gravity data from GRACE (Velicogna et al., 2014) in cm water equivalent overlaid on a reconstruction of ocean temperature at 391 m depth combining observations and ocean modelling from the Southern Ocean State Estimate (SOSE) (Mazloff et al. (2010) updated) showing high mass loss near warm, salty, circumpolar deep water (+2°C). (Right) Time series of mass balance of Antarctica for different time periods since 1979 using the mass budget method (updated from Rignot et al., 2011) with the partitionining of the loss (red) or gain (blue) between surface mass balance processes and ice dynamic processes.

3.2.1.2 *West Antarctic Ice Sheet, Mass Budget Estimates, Altimetry Results, GRACE, Partitioning of the Mass Budget*

The mass balance of West Antarctica is better known than for East Antarctica because the signal is larger and the GIA correction is lower. A study extending over 1979–2016 indicates a mass balance decreasing from –34 ± 9 Gt yr^{-1} in 1979–2003 to –112 ± 12 Gt yr^{-1} in 2003–2016, with most of the loss from the Amundsen Sea Embayment (ASE) which regroups glaciers with a combined sea level potential of 1.2 m (updated from Rignot et al., 2011). Significant dynamic losses are recorded along Getz Ice Shelf, west of

ASE and for GeorgeVI and Stange Ice Shelves, in the Bellingshausen Sea. GRACE indicate a mass loss of 133 ± 18 Gt yr^{-1} for West Antarctic with an acceleration of 11.4 Gt yr^{-2} for the time period 2002–2016 and a loss of 30 ± 9 Gt yr^{-1} for the Peninsula (updated from Velicogna et al. (2014)). A Bayesian method indicates a loss of 112 ± 10 Gt yr^{-1} for 2003–2013 for West Antarctica and 28 ± 7 Gt yr^{-1} for the Peninula (Martín-Español et al., 2016). Over a short time period, Cryosat-2 indicates a mass loss is 134 ± 27 Gt yr^{-1} for 2010 to 2013 for West Antarctica and 23 ± 18 Gt yr^{-1} for the Peninsula (McMillan et al., 2014). Overall, a rapid loss in mass in the ASE sector of West Antarctica and in the Antarctic Peninsula is *virtually certain*, with *high confidence*, with broad agreement on the magnitude and acceleration of the loss from multiple techniques.

On a centennial time scale (1900–2010), firn and ice cores revealed positive accumulation trends in the Peninsula and eastern West Antarctica (Thomas et al., 2013; Wang et al., 2017), negative trends for the western West Antarctica, and no significant trend in the central part (Wang et al., 2017) (*high confidence*).

Anomalies in SMB very likely play a negligible role in the total mass loss of West Antarctica compared with changes in ice discharge, with *high confidence* (Sutterley et al., 2014). In the Peninsula, the long term changes in SMB are smaller than the mass changes caused by glacier dynamics in Graham Land (Pritchard et al., 2012) and George VI (Wouters et al., 2015; Hogg et al., 2017). Mouginot et al. (2014) report a 77% increase in total discharge in the ASE since the 1970's, with rapid increases on the Pine Island, Thwaites and Smith glaciers, along with a grounding line retreat of 1km yr^{-1} for Pine Island and Thwaites and 2 km yr^{-1} for Smith. Rapid grounding line retreat is also pervasive along the Bellingshausen Sea over the past 40 years (Christie et al., 2016; Hogg et al., 2017).

3.2.1.3 Greenland Ice Sheet

Lewis et al. (2017) analyzed 25 NASA Operation IceBridge snow radar flights totaling >17,700 km in 2013 to 2014 to analyze snow accumulation in the dry snow and percolation zones over the past 100–300 years, back to 1712 AD with 10-year intervals in the upper part and 20 years in the lower part. Good agreement was found with overlapping results of Overly et al. (2016) and Karlsson et al. (2016). Significant accumulation trends since 1712 were found neither in the snow radar data nor in the nine annually resolved ice cores that were used to validate the radar data. The absence of an accumulation trend since ~1700 AD is in line with previous assessments (Andersen et al., 2006), which suggest that it is *virtually certain* that there has been no increase in snowfall accumulation in Greenland over the last several centuries, with *high confidence*.

Melt of snow and ice and subsequent runoff *very likely* constitutes a major component of Greenland SMB, with *high confidence*. In summer, melt in the lower ablation zone in the south and west can be continuous (Van den Broeke et al., 2011) with peak daily/seasonal melt rates observed at ~0.3/9 metres of ice in the exceptional melt summer of 2012 (Fausto et al., 2016). Because ablation observations (Machguth et al., 2016a) and automatic weather stations enabling full melt calculations (Citterio et al., 2015; Kuipers Munneke et al., 2018b) are relatively abundant on the GrIS, regional climate models (RCMs) are well constrained since the IGY (1957-58). Gridded SMB fields from RCMs agree well with observations (Lucas-Picher et al., 2012; Vernon et al., 2013; Noël et al., 2015), but marginal outlet glaciers, where runoff is largest, remain poorly resolved. Based on RACMO2.3 11 km resolution output, integrated over the GrIS for the climatological period 1961–1990: melt ≈ 435 Gt yr^{-1}, rainfall ≈ 25 Gt yr^{-1}, refreezing ≈ 200 Gt yr^{-1} and runoff ≈ 260 Gt yr^{-1} (Van den Broeke et al., 2016). When Noël et al. (2016) statistically downscaled RCM output from 11 to 1 km using MODIS ice albedo for narrow and often dark glacier tongues, runoff increased by 30%, which implies that Greenland SMB cannot be accurately determined at typical RCM resolutions of 5–15 km.

In west Greenland (Humphrey et al., 2012) but especially in south-east and northwest Greenland, i.e., areas with high accumulation and high summer melt rates (Kuipers Munneke et al., 2014a), perennial firn aquifers store liquid water year-round (Forster et al., 2013) before it drains in crevasses downslope (Poinar et al., 2017). The observed firn aquifers represent a minimum area of ~18,000 km^2 at an average elevation of ~1600 m asl (Miège et al., 2016) and an water volume of ~140 Gt (Koenig et al., 2014), i.e., similar to the annual runoff from Greenland.

In the early 1990s, the Greenland Ice Sheet appears to have been in a state of near-balance (Hanna et al., 2013; Khan et al., 2015). A significant summer warming of ≈ 2°C since the early 1990s has increased modelled Greenland melt by 35% and runoff even by > 40%, while precipitation and sublimation did not appreciably change (Van den Broeke et al., 2016). The increase in runoff is responsible for most of recent Greenland ice mass loss (Enderlin et al., 2014; Andersen et al., 2015), accounting for 42% between 2000 and 2005, increasing to 64% between 2005 and 2009 and 68% between 2009 and 2012.

In Greenland, the mass loss over a short period from Cryosat-2 for 2011–2016 averaged 269 ± 51 Gt yr^{-1} (McMillan et al., 2016). Similarly, the MBM detects a mass loss of 247 ± 28 Gt yr^{-1} between 2000 and 2012 (Enderlin et al., 2013), with a growing fraction of the loss controlled by surface melt. Using GRACE data, the mass loss averages 271 ± 42 Gt yr^{-1} in 2002–2016, with an acceleration of 10.6 ± 1 Gt yr^{-2} (Velicogna et al., 2014). Over a longer time scale, Rignot et al. (2011) find a mass loss of 103.6 ± 31 Gt yr^{-1} for 1972–2017, increasing from 14.7 ± 33 Gt yr^{-1} in 1972–1990, 35.4 ± 30 Gt yr^{-1} in 1990–2000, to 202.4 ± 30 Gt yr^{-1} in 2000–2010 and 301 ± 30 Gt yr^{-1} in 2010–2017 (Figure 3.3). The mass loss is dominated by the high-accumulation, high discharge regions in north east and south east and north west Greenland (Figures), with a cumulative sea level contribution of about 3mm each since 1972, followed by the northeast and central east, each contributing about 2 mm since 1972. Since the mid 2000s, there has been a marked acceleration of the mass loss as well as a spreading of the loss to the entire ice sheet, including the far north regions. The glacier mass loss is dominated by the evolution of marine-terminating glaciers and a widespread increase in runoff. Overall, Greenland is *virtually certain* to have lost mass since the early 1990s, with *high confidence*.

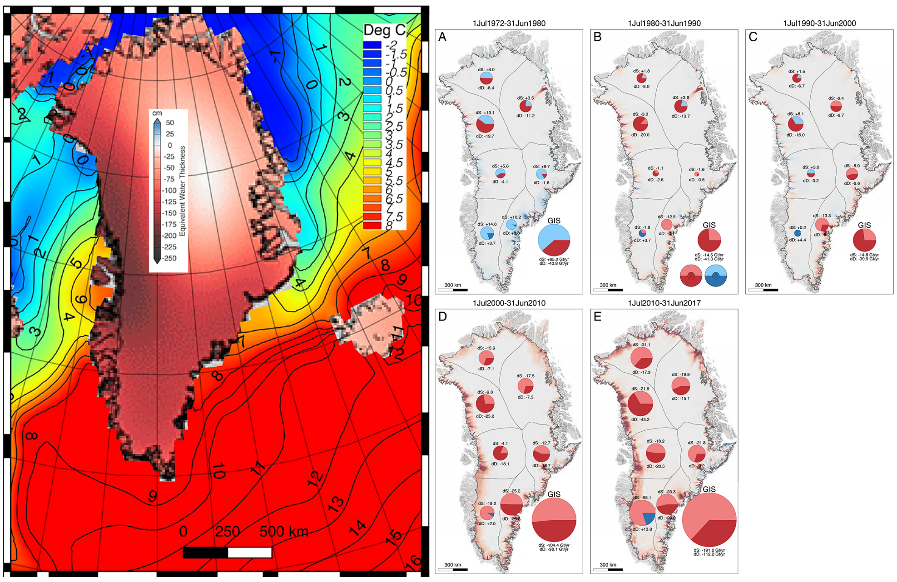


Figure 3.3: (Left) Cumulative mass balance of the Greenland Ice Sheet from the GRACE time-variable gravity data from 2002-2017 (updated Velicogna et al. (2014)) in cm water equivalent overlaid on ocean temperature at 168 m depth combining observations and ocean modelling from the ECCO project and showing the presence of warm, salty water of Atlantic origin along south east Greenland (Forget et al., 2015). (Right) time series of mass balance from the mass budget method since 1972 over different time periods with the partitioning of the loss (red) or gain (blue) between surface mass balance processes and ice dynamics processes (updated Rignot et al., 2011).

3.2.1.4 Polar Glaciers

The mass balance of glaciers including those in the polar regions has been addressed in a comprehensive and uniform manner with GRACE gravity and altimetry data (Jacobs et al., 2011; Gardner et al., 2013). In recent years, better glacier inventories, refined GRACE analysis techniques, and longer time series have improved the results (Schrama et al., 2014; Reager et al., 2016; Rietbroek et al., 2016). In both Alaska (Larsen et al., 2015) and Canada (Millan et al., 2017), mass loss is dominated by SMB processes (94% and 90% respectively), with ice dynamics playing a strong role only in a few glaciers.

[PLACEHOLDER FOR SECOND ORDER DRAFT: text to be updated with forthcoming publications and figure added]

3.2.2 Drivers of Change

3.2.2.1 Ice Sheet Accumulation

Since AR5, we have learned more about the relationship between Antarctic ice sheet accumulation, temperature changes, and variability in large-scale atmospheric circulation. New temperature reconstructions show decadal (1958-2012) warming over West Antarctica and the Peninsula, but no change in East Antarctica (Nicolas and Bromwich, 2014). During the 1990s, West Antarctica experienced record warmth relative to the past 200 years, but similar warmth occurred earlier (Steig et al., 2013). In the Peninsula, strong warming between the mid-1950s and the late 1990s was followed by cooling (Wessem et al., 2017), bringing temperature trends within bounds of natural variability (Turner et al., 2016; Smith and Polvani, 2017). There is now *low confidence* that the observed Antarctic ice sheet accumulation increase since 1800 (Thomas et al., 2017; Medley and Thomas, 2018) is associated with atmospheric warming (Frieler et al., 2015).

It is *likely* that the Southern Hemisphere extratropical circulation has intensified and shifted poleward in austral summer between 1950 and 2012 (Arblaster et al., 2014; Swart et al., 2015), but there is *medium confidence* in these changes given the limited observations and spread in magnitude across various datasets and reanalyses. Paleoclimate reconstructions suggest the austral summer Southern Annuar Mode (SAM) is in its most positive extended state for the past 600 years (Abram et al., 2014; Dätwyler et al., 2017). This intensified atmospheric circulation has likely caused accumulation increases/decreases on the western/eastern side of the Peninsula (1979 to 2013) (Marshall et al., 2017) (*medium confidence*). A deepening of the Amundsen Sea Low over recent decades (Raphael et al. 2015) explains the geographaphically-variable accumulation trends (1900-2010) across West Antarctica (Section 3.2.1.2) (*high confidence*). Emerging drivers of Antarctic accumulation variability (*low confidence*) are katabatic winds that sublimate a significant fraction (17%) of snowfall before it reaches the ground (Grazioli et al., 2017), atmospheric rivers causing regional snowfall anomalies such as in Dronning Maud Land (2009-2011, (Gorodetskaya et al., 2014)), and the impact of sea ice cover on inland precipitation rates (Lenaerts et al., 2016).

Since the early 1990s, summer in west Greenland has warmed by ~2°C and winter by ~ 5°C (Hanna et al., 2012; Box, 2013), but accumulation shows no long-term increase (Section 3.2.1.3), (*high confidence*). The leading atmospheric circulation modes (Section 3.A.1.1) driving Greenland accumulation variability are the Atlantic Multidecadal Oscillation (AMO), possibly due to a warmer North Atlantic during AMO positive conditions, and the wintertime North Atlantic Oscillation (NAO), with greater accumulation in northern/western Greenland during NAO negative conditions, when there is enhanced southerly flow of warm, moist air masses into Baffin Bay (Mernild et al., 2015; Osterberg et al., 2015; Wong et al., 2015). Positive correlations are found between north Greenland ice sheet accumulation and enhanced summertime accumulation, expressed by the Greenland Blocking Index (Hanna et al., 2016). These associations between atmospheric indices and accumulation indicate with *high confidence* that variability of large-scale atmospheric circulation is an important driver of accumulation changes in Greenland.

3.2.2.2 Ice Sheet Surface Melt

Since AR5, both the Antarctic and Greenland ice sheets have experienced exceptional melt events extending to high elevation. In January 2016, melt occurred up to ~1000 m elevation in the Ross Sea sector of West Antarctica, linked to sustained advection of warm and cloudy marine air, *likely* favoured by the strong El Niño (Section 3.A.1.2) (Nicolas et al., 2017) (*low confidence*). In Greenland, increased high-pressure blocking and warm air advection (Hanna et al., 2016; McLeod and Mote, 2016) culminated in an exceptional event in July 2012, when surface melt occurred on the summit of the ice sheet (Nghiem et al., 2012; Tedesco et al., 2013; Hanna et al., 2014).

Observed widespread surface melt and supraglacial meltwater flow at low elevation, including East Antarctica (Langley et al., 2016), are not recent phenomena (Kingslake et al., 2017), (*high confidence*). Since AR5, we have learned that increased föhn winds resulting from a more positive SAM (Cape et al.,

2015) have *likely* caused increased melting on the eastern Peninsula ice shelves (Grosvenor et al., 2014; Luckman et al., 2014; Elvidge et al., 2015). For example, föhn-driven wintertime melt produced ~23% of 2015/16 annual melt on Larsen C ice shelf (Kuipers Munneke et al. 2018). For Greenland, a negative North Atlantic Oscillation (NAO) index explains ~ 70% of summer warming since 2003 ((Fettweis et al., 2013b; Mioduszewski et al., 2016); *medium confidence*).

Reduced snow cover and biological agents (Stibal et al. 2017) caused a −1.2+/−0.9% summer albedo reduction in the ablation area of the Greenland Ice Sheet between 2000−2017 (Tedesco et al. 2015; Box et al. 2016) (*medium confidence*). Clouds increase snow melt by increasing the surface radiation balance (Bennartz et al. 2013; Van Tricht et al. 2016), (*high confidence*), and water vapor transport to Greenland was stronger during 2000-2015 than 1979-1994 (Mattingly et al. 2016). Clear skies drove the recent melt increase in the western ablation zone, where the cloud radiative effect is reversed (Hofer et al. 2017), (*medium confidence*). (Liu et al. 2016) suggests that reduced summer sea ice favours blocking events over Greenland, enhancing melt, but (Stroeve et al. 2017) find only weak correlations between Greenland melt and sea ice (*low confidence*).

3.2.2.3 Ice-Ocean Interactions

Recent studies suggest with *medium confidence* that an anomalous inflow of subtropical waters driven by atmospheric changes, multidecadal natural ocean variability (Andresen et al., 2012), and a long-term increase in the North Atlantic's upper ocean heat content since the 1950s, all contributed to a warming of the subpolar North Atlantic (Häkkinen et al., 2013). However, deep knowledge of the processes by which warmer ocean waters are driven toward the Greenland ice sheet retreat remains elusive (Straneo et al., 2013; Xu et al., 2013b; Bendtsen et al., 2015). Modelling studies indicate with *high confidence* that water temperatures near the grounding zone of Greenland outlet glaciers are critically important to their stability (O'Leary and Christoffersen, 2013). However, no single oceanic or atmospheric trigger for outlet glacier retreat has been identified (Murray et al., 2015; Cowton et al., 2016; Miles et al., 2016). Bed characteristics (Enderlin et al., 2013; Morlighem et al., 2016), and subglacial hydrology (Gladish et al., 2015) also play important roles, providing *low confidence* in linking Greenland outlet glacier behaviour to ocean forcing on seasonal and interannual time scales (Straneo et al., 2016).

Since AR5, continental-scale ice sheet models have been used to simulate the 3D advance and retreat of major outlet glaciers (Todd and Christoffersen, 2014; Morlighem et al., 2016; Muresan et al., 2016; Bondzio et al., 2017). Patterns of acceleration and dynamic thinning in these models is triggered by prescribed changes in terminus position and outlet glacier geometry (Muresan et al., 2016; Bondzio et al., 2017). However, most models fail to account for the effect of submarine melt rates on frontal ablation (Rignot et al., 2010) and the feedback of these melt rates on calving rates (Todd and Christoffersen, 2014; Benn et al., 2017a) (*medium confidence*). Overall there *is low confidence* in the understanding of the response of marine-terminating outlet glaciers to forcing, cautioning that extrapolating behaviour from a small sample of studied glaciers to others is at present unwarranted (Carr et al., 2013).

In Antarctica, the floating ice shelves buttress 90% of the outflow from the ice sheet (Depoorter et al., 2013; Rignot et al., 2014). Since AR5, longer and more detailed records of ice shelf (Paolo et al., 2015; Christianson et al., 2016; Khazendar et al., 2016), and ocean (Jenkins et al., 2016; Webber et al., 2017) change have contributed to *high confidence* that changes in buttressing resulting from ice shelf melting are responsible for most of the inland thinning, although there is *limited evidence* of this effect outside the Amundsen Sea sector (Khazendar et al., 2013; Cook et al., 2016; Rintoul et al., 2016; Walker and Gardner, 2017).

Around most of the coast, near-freezing continental shelf waters shield the ice shelves from warmer Circumpolar Deep Water (CDW) that is found off-shelf (Figure 3.4). There is *high confidence* that a combination of easterly wind stress (Stewart and Thompson, 2015), which drives cold surface waters to the south, and sea ice formation in coastal polynyas (Herraiz-Borreguero et al., 2015; Darelius et al., 2016), that increases the density of the cold waters, keep CDW off the continental shelf. However, there is *low confidence* in the relative importance of those two processes, and *limited evidence* for how changes in either would affect shelf water temperatures. There is *high agreement* that a reduction in sea ice production

(Timmermann and Hellmer, 2013) or a weakening of easterly winds (Spence et al., 2014) would cause warming, but *low confidence* in the thresholds for change to a warmer environment (Hellmer et al., 2017). In the Amundsen and Bellinghausen seas, CDW intrudes onto the continental shelf, driving ice shelf melt rates two orders of magnitude higher than elsewhere. There is *limited evidence* that changes in the thickness of the CDW layer have controlled recent variability in ice shelf melting (Dutrieux et al., 2014). There is *medium confidence* that winds, either directly (Dutrieux et al., 2014; Kimura et al., 2017) or indirectly through their influence on buoyancy forcing in coastal polynyas (St-Laurent et al., 2015; Webber et al., 2017), drive the changes in CDW layer thickness. Winds over the Amundsen Sea are highly variable owing to complex interactions between SAM, ENSO and the Amundsen Sea Low (Turner et al., 2016). There is *limited evidence* that ENSO related variability triggered change on Pine Island Glacier in the 1940s (Smith et al., 2016a).

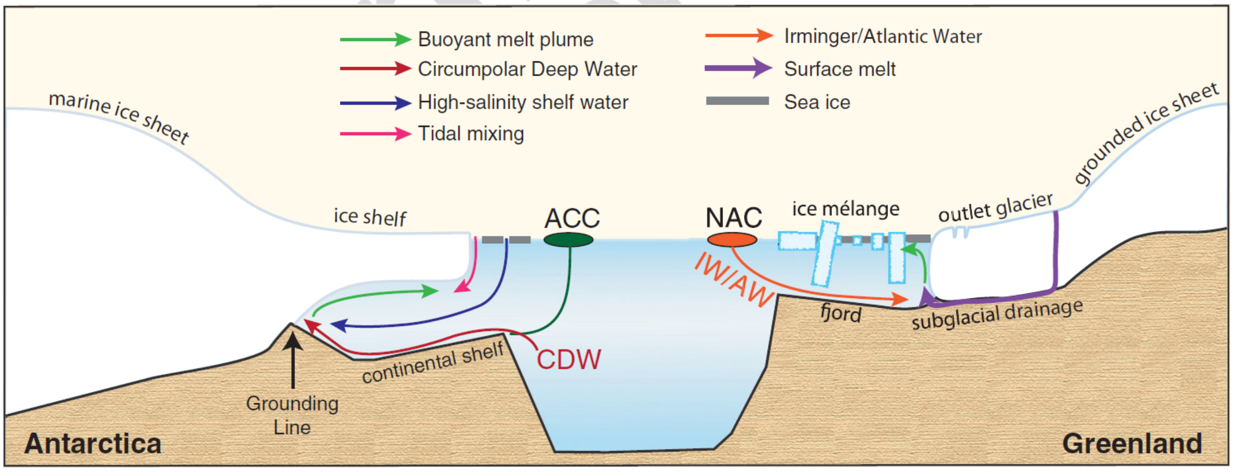


Figure 3.4: Schematic illustration of the ocean processes influencing ice shelves and outlet glaciers (Joughin et al., 2013). Melting beneath Antarctic ice shelves occurs through a combination of three processes (Jacobs et al. 1992). The first occurs where dense, high-salinity shelf water is formed near the ice-shelf front during winter sea-ice growth. Although this water is at the surface freezing point, it can melt ice when it sinks to depths because it is above the local pressure melting point. The second occurs where tidal mixing moves seasonally warmed near-surface ocean water beneath the shelf front. Both of these processes are active for ice shelves with cold cavities. In contrast for warm ice shelf cavities, melting is dominated by the presence of a sub-surface, warm water mass (Circumpolar Deep Water), originating from the Antarctic Circumpolar Current (ACC). Ocean melting of Greenland outlet glaciers is driven by analogous warm waters, namely the Irminger Water (IW) along the western and south-eastern coasts and Atlantic Water (AW) elsewhere. Both of these subsurface water masses originate with the North Atlantic Current (NAC). Where melting occurs, the buoyancy of the resulting meltwater plume produces positive feedback driving further melt, which may be enhanced where subglacial meltwater is present (Jenkins 2011).

3.2.2.4 *Detection and Attribution of Glacier and Ice Sheet Changes*

Recent progress has been made in attributing glacier changes to natural and anthrogenic forcings. Marzeion et al. (2014) leveraged formal attribution techniques, glacier observations, an idealized model of glacier evolution and CMIP5 model historical outputs to attribute negative global-scale glacier mass balance since 1991, equivalent to 30 mm sea level rise, to anthropogenic forcing (*high confidence*). This is consistent with modeling that demonstrates glacier retreat over the last century is a clear indicator of regional climate change (Roe et al., 2016)(*high confidence*).

It is challenging to attribute observed changes in ice sheets to natural and anthropogenic forcings because regional climate variability is an important driver of ice sheet accumulation, melt, and ice-ocean interactions in Greenland and Antarctica (Sections 3.2.2.3 and 3.2.2.4) (Wouters et al., 2013; Turner et al., 2017a). In Greenland, significant natural climate variability including the Atlantic Multidecadal Oscillation (AMO) (Section 3.A.1.1) (Ding et al., 2014; Ding et al., 2017) and other drivers, make it difficult to attribute Greenland mass loss acceleration (Wouters et al., 2013) to anthropogenic climate change (*medium confidence*). Similarly in Antarctica, attribution is challenging because ice-ocean interactions remain poorly understood (Section 3.2.2.3), and atmospheric changes are affected by a combination of greenhouse gas increases, stratospheric ozone depletion (Waugh et al., 2015) and tropical Pacific sea surface conditions (Schneider et al., 2015a; England et al., 2016; Raphael et al., 2016; Clem et al., 2017a) (*medium confidence*).

Current ice-sheet attribution studies generally exclude both ice dynamics and ice-ocean interactions, as in recent studies on source-partitioned sea level rise attribution (Slangen et al., 2016). To circumvent lack of available observations, other studies of ice sheet change have approached identification of a forced signal in ice sheet conditions through 'emergence' approaches, in which a signal is detected directly in model simulation data. In this way, Fyke et al. Fyke et al. (2014b) identified an emerging role for anthropogenic forcing in driving Greenland margin melting and interior snowfall trends, and Previdi and Polvani (2016) similarly suggested that an anthropogenic signal in integrated Antarctic snowfall should emerge by the mid-21st century. Progress in coupled atmosphere-ocean-ice-sheet model development but also in reconstructions of historical ice sheet-relevant climate conditions is required before the formal attribution of ice sheet changes improves.

3.2.3 Rapid and/or Irreversible Changes

3.2.3.1 Ice Shelf Collapse

Mass loss from the Antarctic Ice Sheet occurs primarily through its ice shelves and is controlled by the erosion of ice by basal melting combined with iceberg calving (Liu et al., 2015b). Since AR5 ice shelves continued to retreat including a large iceberg detachment from the Larsen C ice shelf in 2017 (Cook et al., 2012; Hogg and Gudmundsson, 2017). Recent observations have shown that ice shelf thinning associated with ocean-driven increased basal melt can *likely* trigger increased iceberg calving implying that the calving is more sensitive to ocean forcing than expected (Liu et al., 2015b) (*low confidence*). Submarine melt can carve subglacial channels beneath ice shelves (Vaughan et al., 2012) and this may promote rifting and fracturing (Vaughan et al., 2012; Dutrieux et al., 2013). Large basal melt rates may excavate and enlarge wide basal crevasses (Bassis and Ma, 2015), as may be occurring at Pine Island Glacier (Jeong et al., 2016). Negative feedbacks that may increase ice shelf stability, *likely* include marine ice deposition in crevasses within cold ocean cavities (Holland et al., 2009; Jansen et al., 2013; Jordan et al., 2014; McGrath et al., 2014). Since AR5, there is *limited evidence* that surface melt may be a stabilizing factor, if supraglacial rivers remove meltwater before it ponds on the ice shelf surface (Bell et al., 2017; Kingslake et al., 2017). These findings suggest that ocean forcing and surface melt affect ice shelf mechanical stability, but the precise mechanisms remain poorly understood. *Confidence* in attribution of ice-shelf calving events, or projections associated with parameterization of these processes remains *low*.

3.2.3.2 Marine Ice Sheet Instability

Thinning or collapse of buttressing ice shelves trigger grounding-line retreat (Konrad et al., 2018) and, if the glacier bed slopes downward inland, small changes in grounding line position may trigger an irreversible grounding line retreat and accelerating mass loss, a process called Marine Ice Sheet Instability (MISI). Observations since AR5 suggest that this process may be already under way in portions of the Amundsen Sea Embayment, like Pine Island and Thwaites Glaciers (Rignot et al., 2014; Christianson et al., 2016) (*medium confidence*). The ability of models to simulate the processes controlling the marine ice sheet instability has improved since AR5, but significant discrepancies in projections remain. These discrepancies result in a large spread in rates of retreat associated with MISI (Favier et al., 2014; Joughin et al., 2014; Cornford et al., 2015; Ritz et al., 2015; Seroussi et al., 2017).

Another mechanism called the Marine Ice Cliff Instability (MICI) suggests that in the absence of an ice shelf, ice-front retreat into an over-deepening basin could lead to runaway ice cliff failure and ice sheet disintegration (Bassis and Walker, 2012). When this mechanism is parameterized in ice sheet models, collapse of significant portions of the West Antarctic Ice Sheet occur in a few centuries (Pollard et al., 2015; DeConto and Pollard, 2016) (*low confidence*). Limited evidence from paleo-records (Section 3.2.3.5) suggests that parts of Antarctica experienced rapid retreat due to MICI in the recent geological past (*low confidence*).

3.2.3.3 Subglacial Water Discharges

Antarctic subglacial hydrology was not described in AR5. Around 50% of the Antarctic ice sheet bed is wet (Siegert et al., 2017). The basal melting produces ~ 65 Gt yr^{-1} of subglacial water (Pattyn, 2010). Over 400 subglacial lakes exist beneath Antarctic ice sheet with a total volume of water of tens of thousands km^3 (Siegert, 2017). The largest of them, subglacial Lake Vostok, comprises ~6000 km^3 (Popov and Masolov, 2007; Lipenkov et al., 2016). Many lakes are related to each other by system of subglacial channels or rivers with substantial amount of water transported between these 'active lakes' (Fricker et al., 2007; Siegert et al., 2007; Siegert et al., 2016). Subglacial lakes exist under most of Antarctica's fast-flowing ice streams, and subglacial water flow continues to the grounding line (Fricker et al., 2007; Carter and Fricker, 2012; Horgan et al., 2013; Le Brocq, 2013), where it exchanges fresh water and nutrients with the ocean (Section 3.3.1.2.4). A few studies demonstrated possible hydrostatic instability of Antarctic subglacial lakes, including Lake Vostok (Erlingsson, 2006). However, calculations based on more detailed and precise data on the lake morphology showed that the Lake Vostok is *very unlikely* to experience catastrophic discharge (Richter et al., 2014) (*medium confidence*).

3.2.3.4 Evidence of Past Rapid Changes in Ice Sheets from Paleo Records

Progress has been made since AR5 in identifying periods of past rapid ice sheet changes in Antarctica between 18,000 years ago and present. Utilising a geological record of ice-rafted debris from the Scotia Sea, (Weber et al., 2014) showed that parts of the West and East Antarctica *likely* underwent rapid retreat during Meltwater Pulse 1a, ~ 14,600 years ago (*medium confidence*). Johnson et al. (2014) and Wise et al. (2017) showed that Pine Island Glacier experienced rapid thinning and grounding line retreat in the early Holocene, ~ 8,000 years ago (*high confidence*). Evidence of rapid thinning of Antarctic outlet glaciers in the early to mid Holocene has been identified from glaciers flowing into the Ross and Weddell Sea embayments (Jones et al., 2015b; Hein et al., 2016; Spector et al., 2017). It is *very likely* that rapid Antarctic ice sheet changes happened in the past (*high confidence*), which increases confidence that these processes are relevant to projections (Section 3.2.4). These past rapid changes have likely been driven by the incursion of Circumpolar Deep Water onto the Antarctic continental shelf (Hillenbrand et al., 2017), MISI (Jones et al., 2015b) and MICI (Wise et al., 2017) (Section 3.2.3.2).

3.2.4 Projections and Models

3.2.4.1 Greenland

Since AR5, model-based projections of the Greenland ice sheet have focused on inter-comparison exercises (Shannon et al., 2013; Nowicki et al., 2013b; Edwards et al., 2014; Nowicki et al., 2016; Goelzer et al., 2017b) and single model sensitivity analyses (Aschwanden et al., 2013; Seroussi et al., 2013; Aðalgeirsdóttir et al., 2014; Chang et al., 2014; Fyke et al., 2014a; Holschuh et al., 2014; Schlegel et al., 2015; Vizcaino et al., 2015; Mosbeux et al., 2016; Saito et al., 2016; Carr et al., 2017; Peano et al., 2017). These studies have provided systematic assessments of the relative importance of different modelling uncertainties for projections.

Updated projections for Greenland Ice Sheet mass loss by 2100 are mostly at the low end of the AR5 range, or lower (Aðalgeirsdóttir et al., 2014; Fürst et al., 2015; Vizcaino et al., 2015). The range from each study appears to reflect the number of GCMs sampled, which are fewer than the AR5 assessment, and may also be influenced by the implementation of SMB forcing. Peano et al. (2017) find a broader range of results than AR5, but their positive degree day parameterisation of ablation may be too sensitive. These studies use coarse resolution ice-sheet models, which may under-represent marine-terminating glacier responses compared with the flowline modelling on which the AR5 assessment is based (Nick et al., 2013) and may also under-estimate ice-elevation feedbacks. Predictions of the magnitude and timing of ice discharge changes differ but, where assessed, agree that future dynamic mass losses are dominated by changes in surface mass balance (*high confidence*), rather than marine-ice loss, or changes in basal lubrication. Spatial patterns show the greatest decrease in ice thickness in southwest Greenland.

Further limitations and gaps that limit our ability to accurately project the future of the Greenland Ice Sheet (Goelzer et al., 2017a; Shepherd and Nowicki, 2017) include implementation of SMB in climate models

(Broeke et al., 2017) and sub-glacial hydrology in ice sheet models (Bueler and van Pelt, 2015; Aschwanden et al., 2016), missing processes of ocean circulation and glacier calving, lack of observations of fjord geometries and basal melting (Section 3.2.2.3); and model initialisation methods (Goelzer et al., 2017a). For stand-alone ice-sheet models, they include implementation of SMB forcing and feedbacks, parameterisation of ocean feedbacks, and for coupled models the conservation of quantities, mismatching resolutions, the treatment of surface albedo, and surface mass balance downscaling.

3.2.4.2 Antarctica

Antarctic ice sheet modelling is a rapidly evolving field. A key area is instability of the West Antarctic Ice Sheet; in particular, if and when a threshold for collapse might be exceeded. Since AR5, regional-scale ice sheet modelling suggests MISI (Section 3.2.3.2) in the Amundsen Sea Embayment is driven by ocean-forced basal melting (Favier et al., 2014; Joughin et al., 2014). This instability may begin elsewhere if basal melting increases or ice shelves collapse.

Continental-scale projections agree the Amundsen Sea Embayment will continue to show the largest grounding line retreat and thinning during the next two centuries (Cornford et al., 2015; Golledge et al., 2015; Ritz et al., 2015; DeConto and Pollard, 2016). Smaller changes are projected from the 22nd century onwards for the Aurora and Wilkes Basins, and for glaciers feeding into the Ronne-Filchner and Ross ice shelves (*low confidence*). Studies disagree whether Thwaites Glacier or the Peninsula are vulnerable in the next one or two centuries. The degree with which MISI or MICI may be self-sustaining if basal melting decreases is unknown (Favier et al., 2014; Joughin et al., 2014; Seroussi et al., 2014a; Feldmann and Levermann, 2015; Arthern and Williams, 2017). Models predict the eventual emergence of connected ocean channels between the Amundsen, Ross and Weddell Seas, first isolating the Peninsula, and ultimately connecting the Amundsen and Ross Seas. Similar retreats in the Aurora and Wilkes Basins are predicted over centuries to millennia (Mengel and Levermann, 2014; Golledge et al., 2015; DeConto and Pollard, 2016; Pattyn et al., 2017). There is *low confidence* regarding the timing of such a deglaciation. There is *low confidence* in predicting the triggers of dynamical change (Sections 3.2.2.2 and 3.2.2.3) (Pattyn et al., 2017; Shepherd and Nowicki, 2017). Kuipers Munneke et al. (2014b) predict vulnerability of Peninsula shelves to warming late this century, but for most others not until next century, later than DeConto and Pollard (2016). Basal melting depends on ocean circulation, for which *confidence* in model projections is *low*. Coupled model studies suggest commonly-used melt parameterisations over-estimate future increases (De Rydt and Gudmundsson, 2016; Seroussi et al., 2017), reducing confidence in current ice-sheet model projections.

Most projections find a threshold response for West Antarctica collapse (*medium confidence*). This might be avoided if basal melting in the Amundsen Sea Embayment remains close to, or lower than, current levels (Joughin et al., 2014; Feldmann and Levermann, 2015; Pattyn, 2017), under RCP2.6 (Golledge et al., 2015; DeConto and Pollard, 2016); or future cumulative carbon emissions less than 600 GtC (Winkelmann et al., 2015), corresponding to around 2°C warming, though Levermann et al. (2013) predict collapse under 1°C warming. There is *very low confidence* in the magnitude of these thresholds due to limited exploration of model uncertainties.

Key sources of uncertainty in projections are model resolution, physical approximations and initial conditions (e.g., Pattyn and Durand, 2013; Nowicki et al., 2013a; Feldmann et al., 2014; Seroussi et al., 2014b; Cornford et al., 2015; Cornford et al., 2016; Nias et al., 2016; Edwards et al., in review)

3.2.4.3 Polar Glaciers

Only two studies since AR5 provide projections for polar glaciers including those surrounding the Greenland and Antarctic ice sheets in response to RCP scenarios. These updated projections (Radić et al., 2014; Huss and Hock, 2015) indicate with *high confidence* that polar glaciers will continue to lose mass in the 21st Century and beyond, although regional differences are apparent. Glaciers in Iceland, Svalbard, Western Canada, and in the Russian Arctic are projected to lose half (RCP2.6) to nearly all (RCP8.5) of their remaining ice by 2100. In contrast, glaciers in Arctic Canada may be less sensitive and/or may experience less warming, and are currently projected to lose ~15% (RCP2.6) to ~50% (RCP8.5) of their mass by 2100 (*medium confidence*).

A limitation of polar glacier projections is that ice dynamics are treated in an idealised manner. Future work should attempt to utlise dynamic glacier models (e.g. Clarke et al., 2015), that also include ablation processes at marine-terminating glacier fronts (Huss and Hock, 2015; McNabb et al., 2015), and the insulating effects of surface debris cover. Model limitations mean that we have *medium confidence* in glacier projections published since AR5.

3.3 Implications of Climate Change for Polar Oceans and Sea Ice: Feedbacks and Consequences for Ecological and Social Systems

3.3.1 Observed Changes in Ocean and Sea Ice

3.3.1.1 Sea Ice

Sea ice insulates the ocean from the atmosphere, provides an essential habitat for mammals, influences navigation and access to the polar regions, and is of high importance to the traditional lifestyle of northern communities. The characteristics of sea ice cover differs between the Arctic and Antarctic. The central Arctic ocean is surrounded by land, and ice circulates within this basin, forced largely by atmospheric circulation. Depending on dynamic and thermodynamic conditions, some Arctic sea ice survives the summer melt season to form multi-year ice. Conversely, the Antarctic continent is surrounded by sea ice, which interacts with the Southern Ocean. Nearly all Antarctic sea ice is seasonal, melting completely during the austral summer. While the primary sources of information (largely satellite data) are the same, the differing physical processes and seasonal regimes of Arctic versus Antarctic sea ice mean observed trends and climate model performance are quite different between the two polar regions. Knowledge on Arctic sea ice includes rich and pervasive Indigenous and Local Knowledge (ILK), primarily from communities across Alaska and the Canadian Arctic.

3.3.1.1.1 Extent and concentration

The pan-Arctic loss of sea ice cover is a prominent indicator of climate change (Figure 3.5). Nearly four decades of consistent satellite observations have documented pronounced declines in sea ice extent (the total area of the Arctic with at least 15% sea ice concentration) for each month of the year (Serreze and Stroeve, 2015; Stroeve and Notz, 2015) (see also Figure 3.6). Changes are largest in summer and smallest in winter, with September trends (month with the lowest sea ice cover; 1979 to 2017) of –83,000 $km^2 yr^{-1}$ (-13.0% per decade), and –41,000 $km^2 yr^{-1}$ (–2.7% per decade) for March (month with the greatest sea ice cover; (Onarheim et al., 2018) (*very high confidence*). Spatially, the regions of summer ice loss are dominated by changes in the East Siberian Sea (explains 22% of the September trend), with large declines also observed in the Beaufort, Chukchi, Laptev and Kara seas (Onarheim and Årthun, 2017). Winter ice loss is dominated by reductions within the Barents Sea, responsible for 27% of the pan-Arctic March sea ice trends (Onarheim and Årthun, 2017). Reconstructions of the sea ice cover back to 1850 using earlier satellite observations, ship and aircraft observations, ice charts, and whaling records shows that Arctic ice loss over the past 2 decades is *likely* unprecedented in at least 150 years (Walsh et al., 2017).

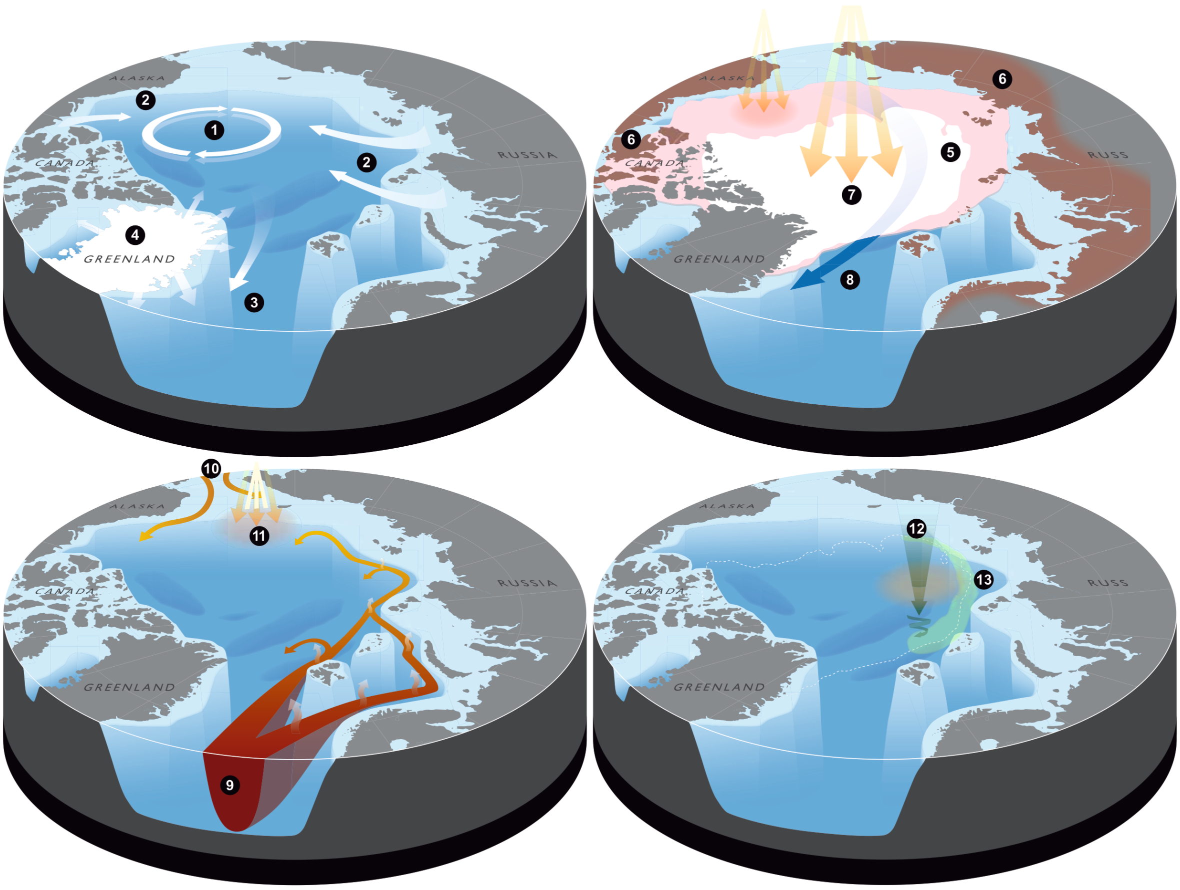


Figure 3.5: Schematic of some of the major Arctic changes assessed in this section. (1) strengthening of the circulation of the Beaufort Gyre {3.3.1.3.1}; (2) increasing discharge of freshwater from rivers to the Arctic Ocean {3.3.1.2.2}; strengthening efflux to lower latitudes through Fram Strait {3.3.1.3.1}; (4) increasing glacial loss from Greenland {3.2.1.3}; (5) retreat of sea ice {3.3.1.1.1}; (6) retreat of seasonal snow cover on land {3.4.1.1.1}; (7) changing ice-albedo feedback {3.A.1.2}; (8) strengthening transport within the Transpolar Drift {3.3.1.1.4}; (9) increasing oceanic heat transport from North Atlantic {3.3.1.2.1}; (10) increasing oceanic heat transport from North Pacific {3.3.1.2.1}; (11) heating of surface layers via insolation {3.3.1.2.1}; (12) carbon drawdown from atmosphere {3.3.1.2.4}; and (13) increasing primary production associated with areas of ice retreat {3.4.4.1.1}.

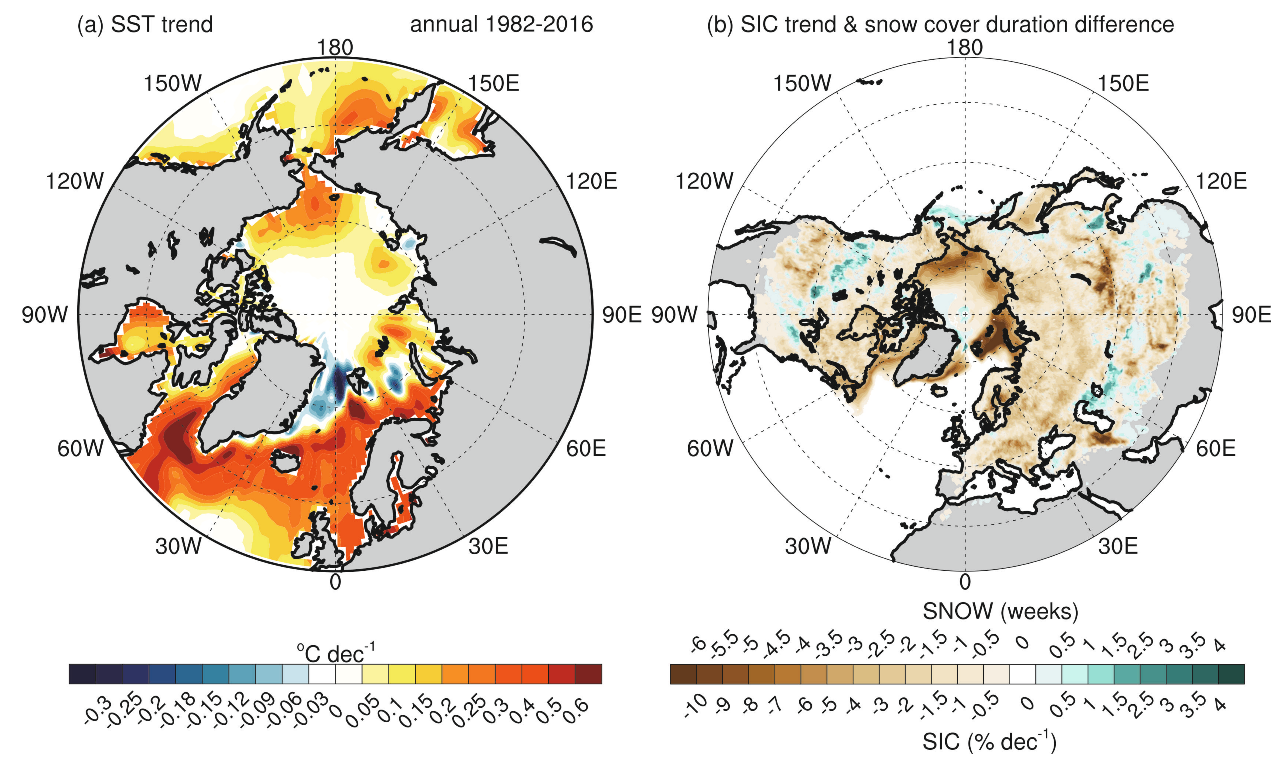


Figure 3.6: (a) Linear trends of annual-mean sea surface temperature for 1982−2016. (b) Linear trends of annual-mean sea ice concentration for 1982−2016, alongside the difference in climatological snow cover duration (in weeks) between the 2006−2015 period and the 1981−1990 period. (a) is from the NOAA Optimum Interpolation Sea Surface Temperature dataset (version 2; Reynolds et al. (2002); https://www.ncdc.noaa.gov/oisst). Sea ice data in (b) were from NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, Version 3 (https://nsidc.org/data/g02202). Snow cover duration (b) was derived from a blend of 4 independent datasets, each covering the 1981-2015 period (Brown et al., 2003; Takala et al., 2011; Brun et al., 2012; Reichle et al., 2017).

The observed reductions in Arctic sea ice cover are strongly linked to warming from increasing concentrations of atmospheric greenhouse gases, with approximately 50 to 60% of the observed sea ice loss driven by external forcing, and the remainder from natural climate variability (Kay et al., 2011; Notz and Marotzke, 2012; Stroeve et al., 2012b; Stroeve and Notz, 2015; Notz and Stroeve, 2016b) (*very high confidence*). While strong anticyclonic circulation over the Arctic Ocean and Greenland in summer have certainly played a role in observed summer sea ice reductions (Ding et al., 2017), anomalous sea ice minima in September are preceded by a wide range of summer atmospheric circulation patterns (Serreze et al., 2016). There is *very high confidence* that the ice-albedo feedback plays a key role in the evolution of summer sea ice cover (Schröder et al., 2014). Earlier melt onset, driven by warm and moist air advection (Mortin et al., 2016) allows for earlier formation of melt ponds (Perovich and Polashenski, 2012) and open water areas (Stroeve et al., 2016). This enhances the ice-albedo feedback, leading to more ice melt in summer (Stroeve et al., 2014a) and thinner ice. Once air temperatures drop below freezing, thermodynamic ice growth is enhanced for thin ice over formerly open water; later freeze-up in autumn means snowfall accumulation on sea ice is delayed, leading to a thinner snowpack (and hence increased thermodynamic ice growth). There is *high confidence* that these two negative feedbacks help to stabilize seasonal sea ice formation, mitigating sudden and irreversible sea ice loss (Stroeve and Notz, 2015).

Coupled climate models show that anthropogenic warming at the Antarctic surface is delayed by the Southern Ocean circulation, which transports heat downwards into the deep ocean (*high confidence*) (Armour et al., 2016). This overturning circulation may explain the weak response of Antarctic sea ice cover to increased atmospheric greenhouse gas concentrations compared to the Arctic. Antarctic sea ice extent has increased overall during the satellite era (since 1979), at an annual-mean rate of $20.2 \pm 4.0 \times 10^3$ km^2 yr^{-1} (*high confidence*) (Comiso et al., 2017), but with a very sharp decline since 2016 (Turner et al., 2017b). The overall increase is composed of near-compensating regional changes, with rapid ice loss in the Amundsen and Bellingshausen seas outweighed by rapid ice gain in the Weddell and Ross seas (Holland, 2014) (Figure 3.7). The regional trends are also strongly seasonal in character (Holland, 2014); only the western Ross Sea has a trend that is statistically-significant in all seasons. The overall trend in ice extent is greatest in autumn, at $26.4 \pm 7.3 \times 10^3$ km^2 yr^{-1} (*high confidence*) (Comiso et al., 2017).

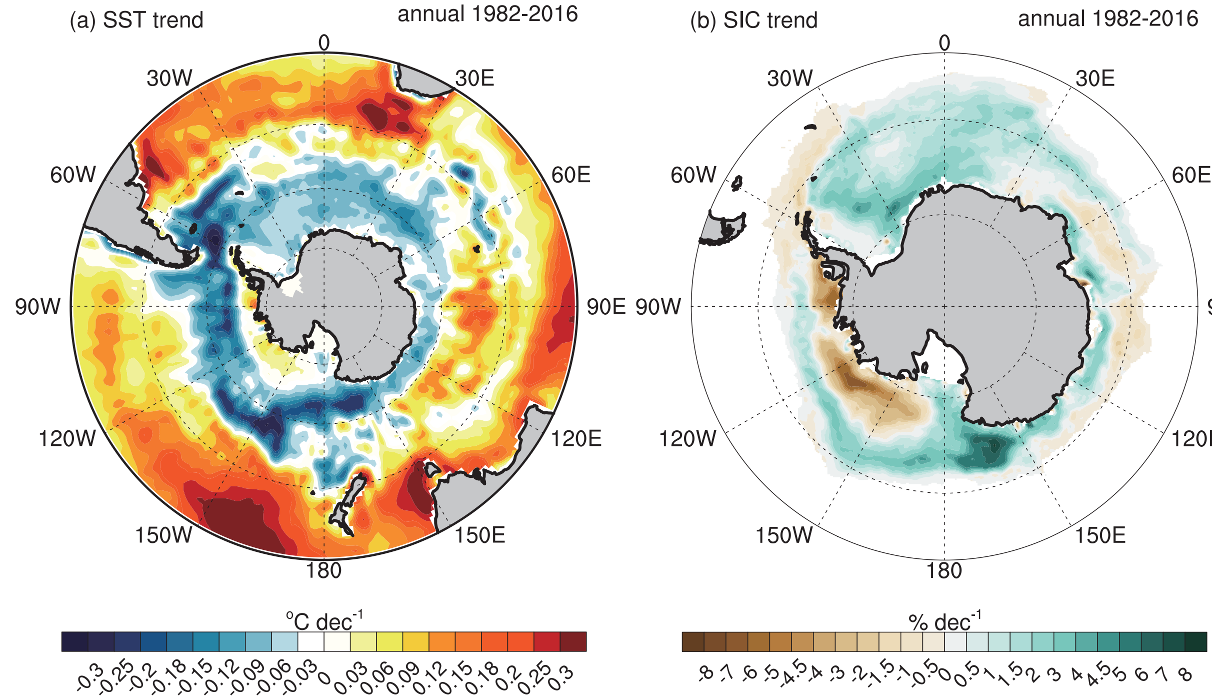


Figure 3.7: (a) Linear trends of annual-mean sea surface temperature for 1982–2016. (b) Linear trends of annual-mean sea ice concentration for 1982–2016. (a) is from the NOAA Optimum Interpolation Sea Surface Temperature dataset (version 2; Reynolds et al. (2002); https://www.ncdc.noaa.gov/oisst). (b) is from the NOAA/NSIDC Climate Data Record of Passive Microwave Sea Ice Concentration, Version 3 (https://nsidc.org/data/g02202).

The regional pattern of observed Antarctic sea ice trends is closely related to meridional wind trends (Holland and Kwok, 2012; Haumann et al., 2014). Poleward wind trends in the Bellingshausen Sea keep sea ice close to the coast (Holland and Kwok, 2012) and advect warm air to the sea ice zone (Kusahara et al., 2017), the reverse being true in the Ross Sea. Wind trends in west Antarctica are strongly affected by tropical Pacific variability (*high confidence*) (Simpkins et al., 2014; Meehl et al., 2016; Purich et al., 2016b), but it is not clear to what extent tropical teleconnections explain the large sea ice increase in the western Ross Sea (Coggins and McDonald, 2015). High latitude atmospheric modes of variability are also influential in west Antarctic sea ice trends, (e.g. the Southern Annular Mode; Appendix 3.A.1.3); in other sectors the wider climate linkages are unclear (e.g., Matear et al., 2015). Ocean-sea ice feedbacks occur as freshwater fluxes during sea ice melt and freeze affect vertical heat flux from the ocean, which may prolong sea ice anomalies (Goosse and Zunz, 2014). While the magnitude of these feedbacks is sufficient to explain the expansion of Ross sea ice (Lecomte et al., 2017), there is *low confidence* in their overall importance.

Ozone depletion in the Southern Hemisphere is linked to strengthening circumpolar westerly winds and the SAM (e.g., Gillett et al., 2013; Christidis and Stott, 2015), which have the potential to affect zonal-mean Antarctic sea ice on two time scales (*medium confidence*) (Ferreira et al., 2015), with an initial sea ice expansion followed by a delayed sea ice decrease. The longevity of the initial ice expansion phase is highly uncertain, as is the magnitude of its effect (Holland et al., 2017). Ozone depletion may also affect meridional winds (Fogt and Zbacnik, 2014; England et al., 2016), but there is *low confidence* that this explains observed sea ice trends (Landrum et al., 2017). It has been suggested that the Antarctic sea ice expansion is caused by the increased freshwater flux from the Antarctic Ice Sheet and ice shelves (Bintanja et al., 2013). Most studies conclude that glacial freshwater input is insufficient to cause a significant expansion (Swart and Fyfe, 2013; Pauling et al., 2017), and there is *medium confidence* that historical sea ice trends have not been driven by glacial meltwater.

Information on Antarctic sea ice changes prior to the modern satellite record is sparse, with *low confidence* in proxy reconstructions of 20th century sea ice variability (Ackley et al., 2003; Hobbs et al., 2016a; Hobbs et al., 2016b).

3.3.1.1.2 Thickness and age

Considerable effort has gone into combining data from multiple satellite altimeter missions to assess ice thickness changes (Kwok et al., 2009; Laxon et al., 2013). These data records show declines in Arctic Basin ice thickness from 2000 to 2012 of –0.58 ± 0.07 m per decade (*high confidence*) (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 (*high confidence*) (Lindsay and Schweiger, 2015). The long-term thinning of Arctic sea ice cover has made it more vulnerable to anomalous atmospheric forcing (as seen in 2007 and 2012) (Stroeve et al., 2008; Zhang et al., 2013) with an overall shorter residence time of sea ice within the Arctic Basin: since 1979 the proportion of ice at least 5 years old declined from 30% to less than 5% (Maslanik et al., 2011; Stroeve et al., 2012a). First-year ice now makes up to 60–70% of the Arctic Basin, compared to only 40% in the early and mid 1980s. Since first-year ice typically grows to 1.5 to 2 m over a winter ice growth season, long-term declines in thickness largely reflect the loss of the perennial ice cover.

In-situ observations of Antarctic sea ice thickness are extremely sparse (Worby et al., 2008). There are no consistent long-term observations from which trends in ice volume are derived. Calibrated model simulations suggest that ice thickness trends follow those of ice concentration, with an overall increase in Antarctic sea ice volume of approximately 30 km^3/y during 1992–2010 (*low confidence*) (Massonnet et al., 2013; Holland et al., 2014).

3.3.1.1.3 Phenology

The Arctic sea ice melt season (onset of liquid water within the snowpack) has extended by more than 10.0 days per decade (*very high confidence*), largely as a result of later freeze-up (+7.5 days per decade), and to a lesser extent earlier melt onset (Stroeve et al., 2014a). The largest trends towards longer open water periods are found in the Barents Sea (+21.8 days per decade) and Chukchi Sea (+16.8 days per decade). Earlier melt onset and later freeze-up both play a role in the Barents Sea (–8.2 and +13.6 days per decade, respectively), whereas the lengthening in the open water season in the Chukchi Sea is largely driven by later autumn freeze-up (+14.1 days per decade). While melt onset trends are generally smaller, they play a large role in the earlier development of open water (Stroeve et al., 2012a; Stroeve et al., 2016), and melt pond development (Perovich and Polashenski, 2012), enhancing the ice-albedo feedback (Perovich et al., 2011; Stroeve et al., 2014a). The timing and magnitude of spring melt pond coverage is also a predictor of the September ice extent (Schröder et al., 2014). Observed reductions in the length of seasonal sea ice cover are reflected in community-based observations of decreased length of time in which activities can safely take place on sea ice (Laidler et al., 2010; Eisner et al., 2013; Fall et al., 2013; Ignatowski and Rosales, 2013).

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

3.3.1.1.4 Motion

Winds associated with the climatological Arctic sea level pressure pattern drive the Beaufort Gyre and the Transpolar Drift Stream, which sequester ice within the central Arctic Basin and export ice out of Fram Strait, respectively, with inter-annual variability in atmospheric circulation strongly influencing ice export (Smedsrud et al., 2011; Smedsrud et al., 2017). As Arctic ice cover has thinned, it is *virtually certain* that drift speeds have increased, both within the Arctic Basin and through Fram Strait (Rampal et al., 2009). While ice export through Fram Strait ranges on the order of 600,000 to 1 million km^2 of ice annually, (approximately 10% of the ice within the Arctic Basin; Smedsrud et al. (2017)), there is only *medium confidence* in observed trends through Fram Strait because of different trends reported from different datasets over non-standard time periods (Kwok et al., 2013; Krumpen et al., 2016; Smedsrud et al., 2017). Advancements in ice tracking algorithms and enhanced satellite radar coverage of the Arctic have supported new understanding of processes governing regional ice fluxes (Howell et al., 2016).

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. Qualitatively, these ice drift trends are supported by agreement with surface wind trends from reanalysis (Holland and Kwok, 2012), but there is only *medium confidence* in these trends due to a small number of ice drift data products derived from a temporally inconsistent satellite record (Haumann et al., 2016).

3.3.1.1.5 Landfast ice

Immobile sea ice anchored to land is referred to as 'landfast'. Long term records of Antarctic fast ice are limited in space and time (Stammerjohn and Maksym, 2017), with a high degree of regional variability in reported trends (Fraser et al., 2011). Very few long-term records of Arctic landfast ice thickness exist, but all exhibit thinning trends in springtime maximum ice thickness. Since the mid-1960s, reported declines are 11 cm per decade in the Barents Sea (Gerland et al., 2008), 3.3 cm per decade along the Siberian Coast (Polyakov et al., 2010), and 3.5 cm per decade in the Canadian Arctic Archipelago (Howell et al., 2016). Over a shorter 1976 to 2007 period, landfast 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 (25% 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 ice duration (~1 week per decade) since the 1970s (Yu et al., 2013; Mahoney et al., 2014). While most Arctic landfast ice melts completely each summer, perrenial landfast ice (also termed an 'ice-plug') occurs in Nansen Sound and the Sverdrup Channel in the Canadian Arctic Archipelago. These ice-plugs were in place continuously from the advent of observations in the early 1960s, until they were both removed during the anomalously warm summer of 1998, and they have rarely re-formed since 2005 (Pope et al., 2017). The loss of this perennial ice is associated with reduced landfast ice duration in the northern Canadian Arctic Archipelago (Galley et al., 2012; Yu et al., 2013) and increased inflow of multi-year ice from the Arctic Ocean into the northern Canadian Arctic Archipelago (Howell et al., 2013).

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

3.3.1.2 Ocean Properties

The Polar Oceans are amongst the most rapidly-changing regions of the world's ocean, with consequences for the storage and cycling of heat, carbon, freshwater and other climatically- and ecologically-important properties. For example, in coupled climate models possessing a carbon cycle, the Southern Ocean accounts for 75 ± 6–22% of the total global ocean heat uptake and 43% ± 3% of carbon (Frölicher et al., 2015) (Appendix 3.A, Figure 2). They are also amongst the most challenging environments from which to obtain data; nonetheless, they are key regions that require understanding and reliable prediction, due to their profound global influence.

3.3.1.2.1 Temperature

AR5 reported that Arctic surface waters warmed from 1993 to 2007, and observations over 1950–2010 show the Atlantic Water Layer warming starting in the 1970s. Warming trends have continued: August linear trends for 1982–2017 reveal summer mixed-layer temperatures increasing at about 0.5°C per decade over large sectors of the Arctic basin that are ice-free in summer (Timmermans et al., 2017) (see also Figure 3.6). This is largely the result of increased solar warming that accompanies sea-ice loss (Perovich, 2016). Between 1979 and 2011, the decrease in Arctic Ocean albedo corresponded to 6.4 ± 0.9 W/m^2 more solar energy input to the ocean (*virtually certain*) (Pistone et al., 2014). This excess solar heat likely mitigates the growth of sea ice by up to 25% in both the Eurasian and Canadian basins (Timmermans, 2015; Ivanov et al., 2016).

While Atlantic Water Layer temperatures have stabilized since 2008, the total heat content in this layer continues to increase, likely due to increased volume inflows (Polyakov et al., 2017). Recent changes have been referred to as the "Atlantification" of the Eurasian Basin; changes are characterized by weaker stratification and enhanced upward heat fluxes further northeast. Polyakov et al. (2017) estimate 2 to 4 times larger heat fluxes in 2014-2015 compared with 2007-2008 (*medium confidence*), with the excess heat likely explaining about 18-40 cm of sea-ice loss in the Eurasian Basin. In the Canadian Basin, the maximum temperature of the Pacific Water Layer increased by about 0.5°C between 2009 and 2013 (*medium confidence*); associated with this was a doubling in integrated heat content (Timmermans et al., 2014). Over 2001-2014, heat transport associated with Bering Strait inflow increased by 60%, due to increases in both volume flux and temperature (*medium confidence*) (Woodgate et al., 2015; Woodgate, 2017).

During 2006–2013, the Southern Ocean accounted for 67–98% of total heat gain in the upper 2000 m of the global ocean (Roemmich et al., 2015)(*high confidence)*. Southern Ocean heating is strongest in the upper 2000 m (Figure 3.8), and peaks in the latitude range 40°S–50°S (Armour et al., 2016) (see also Appenix 3.A, Figures 2 and 3). This contrasts with the surface waters south of the core of the Antarctic Circumpolar Current, which have warmed on average only by 0.02°C per decade, relative to a global SST trend of 0.08°C per decade since 1950 (Armour et al., 2016) (*high confidence),* and which have exhibited cooling in more recent decades (see also Figure 3.7). There is *high confidence* that the observed pattern of upper-layer warming is driven by the upper cell circulation, whereby heat uptake at the surface by newly-upwelled waters is transmitted to the ocean interior in intermediate depth layers (Armour et al., 2016). The warming on the northern side of the ACC associated with this pattern appears too deep to be caused trivially by air-sea fluxes however (Gille, 2014); instead, heave (vertical movement of density surfaces) is more important (*medium confidence)* (Desbruyeres et al., 2017; Gao et al., 2018). Below the surface south of the ACC, warming extends close to the Antarctic continent, particularly on the shelf along the Amundsen-Bellingshausen Sea where increases of 0.03°C yr^{-1} have been observed between 1975–2012 (Schmidtko et al., 2014) (*medium confidence*; see also Section 3.2.2.3).

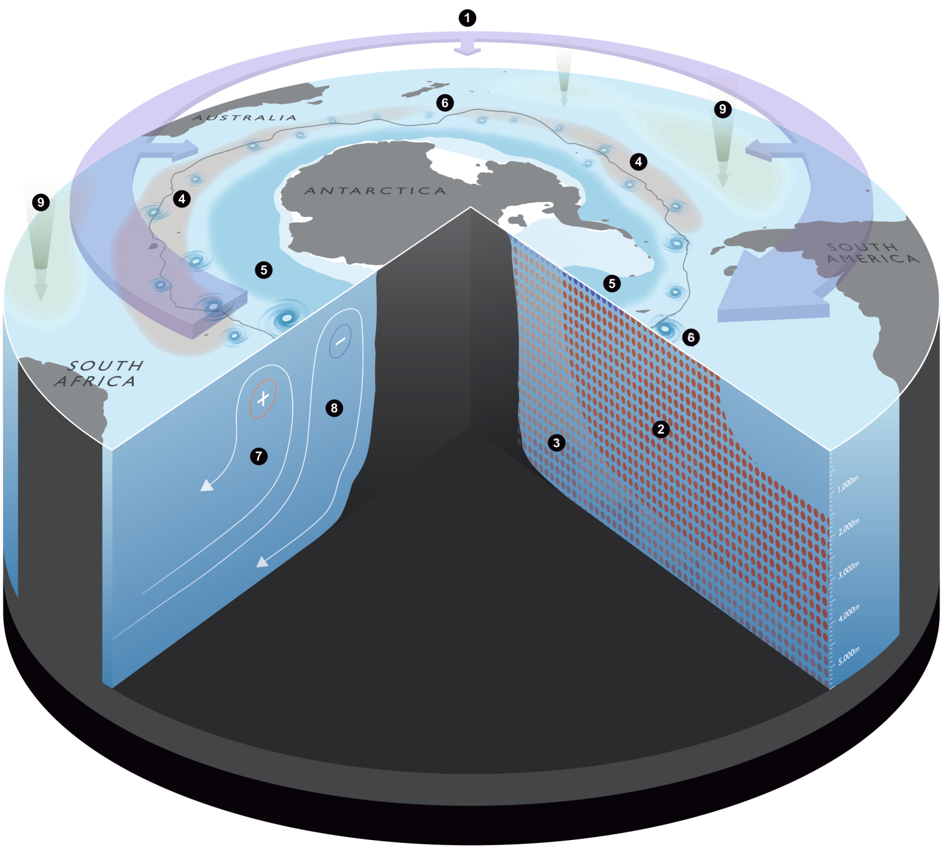


Figure 3.8: Schematic of some of the major observed Southern Ocean changes discussed in this section.
1. Increased strength and poleward contraction of circumpolar winds (3.A.1.3)
2. Strong warming in upper and mid-depth interior of the ocean (3.3.1.2.1)
3. Significant warming and freshening in deep and abyssal ocean (3.3.1.2.1; 3.3.1.2.2)

4. Warming of surface waters toward northern part of circumpolar Southern Ocean (3.3.1.2.1)
5. Freshening and delayed warming of surface layers in southern part of circumpolar Southern Ocean (3.3.1.2.1; 3.3.1.2.2)
6. Increased intensity of Southern Ocean eddy field (3.3.1.3.2)
7. Strengthening of upper cell of overturning circulation (3.3.1.3.3)
8. Reduction in export of deep and abyssal waters from Southern Ocean (3.3.1.3.3)
9. Increased carbon drawdown from the atmosphere and ocean acidification (3.3.1.2.4)

Globally, around 19% of the excess anthropogenic heat in the Earth system is stored in the ocean beneath 2000 m, with the largest part of this (6% of global total heat excess) located in the deep Southern Ocean south of 30°S (Frölicher et al., 2015; Talley et al., 2016). The AR5-quantified warming of these waters was recently updated (Desbruyeres et al., 2017) to an equivalent heat uptake of 0.07 ± 0.06 W m^{-2} below 2000 m since the beginning of the century, resulting in an extra 34 ± 14 TW south of 30°S from 1980–2012 (Purkey and Johnson, 2013) *(medium confidence)*. Isopycnal heave has been identified as the main factor influencing Antarctic Bottom Water (AABW) properties away from Antarctica, while the loss of AABW in the Indian and Pacific basins close to the Antarctic continent is consistent with a warming and freshening of these waters (Purkey and Johnson (2013); see also Chapter 5).

3.3.1.2.2 Salinity

Salinity is the dominant variable that determines density in the polar oceans; as such it exerts major controls on stratification, circulation and mixing. Changes in salinity are induced by changes in freshwater discharged to the ocean, with the potential to impact water mass formation and overturning circulation over large spatial scales. Changes in freshwater discharge to polar waters have been invoked as having the potential to inhibit deep convection, and to weaken or disrupt the Atlantic Meridional Overturning Circulation (e.g. Thornalley et al. (2018); see Chapter 6).

Following increases of Arctic Ocean freshwater content reported in AR5, recent Arctic-wide estimates yield a freshwater increase of 600 ± 300 km^3/yr over 1992 to 2012, with about 2/3 of this trend attributed to a decrease in salinity, and the remainder to a thickening of the freshwater layer (Rabe et al., 2014; Haine et al., 2015a; Carmack et al., 2016). The Beaufort Gyre region has seen an increase in freshwater (*medium confidence*) of about 40% (6,600 km^3) over 2003-2017, with total freshwater content in the region reaching 23,500 km^3 in 2017 (Krishfield et al., 2014; Proshutinsky et al., 2015). The strengthening and freshwater accumulation of the Beaufort Gyre is very likely the result of a strong dominance of clockwise wind patterns over the Canadian Basin between 1997 and 2016, in combination with freshwater input 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 with *low* to *medium* confidence to be about 180 km^3 between 2003 and 2014 (Armitage et al., 2016). 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 through Bering Strait in 2014 of around 3,500 km^3 (*medium confidence)* (Woodgate, 2017). Freshwater flux from rivers is also increasing (Section 3.4.1.2.2).

Observed salinity trends in the Southern Ocean are consistent with those reported in AR5; subsequent studies have increased our confidence in their magnitude and sign, although sparse and short records still represent a major source of uncertainty. Multi-decadal salinity change over 1950–2010 show a persistent freshening of surface waters over the whole Southern Ocean, with trends 0.01–0.05 psu/60 years in mode and intermediate waters to below 1500 m (Skliris et al., 2014). Averaged circumpolarly, de Lavergne et al. (2014) observe a freshening south of the ACC of 0.0011 ± 0.0004 psu yr^{-1} in the upper 100 m since the 1960s. This trend intensifies over the Antarctic shelves and freshening of up to to 0.01 psu yr^{-1} is observed over much of the shelf, except along the western Antarctic peninsula (Schmidtko et al., 2014). Recently, there has been increased recognition of the importance of sea ice in driving Southern Ocean salinity changes; Haumann et al. (2016) demonstrate that wind-driven sea ice export has increased by 20 ± 10 Sv from 1982–2008, and that this may have driven freshening of 0.002 ± 0.001 yr^{-1} in the surface and intermediate waters (*medium confidence*).

For Antarctica, there is *limited evidence* for both an increase in snowfall in most coastal regions over the past 200 years (Thomas et al., 2017). Freshwater input from the ice sheet is divided approximately equally between calving of icebergs and melting of contiguous 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 sea since the 1990s, but *low confidence* on trends in other sectors (Paolo et al., 2015). There is *low confidence* on freshwater inputs from iceberg melting, because calving rates are naturally highly variable and difficult to quantify (Liu et al., 2015b), while much of the meltwater input occurs far from the site of calving (Merino et al., 2016).

Based on repeat hydrographic profiles between 1980–2012, Purkey and Johnson (2013) show that Antarctic Bottom Water (AABW) has freshened by 0.0001–0.0005 psu yr^{-1} in most Southern Ocean basins and up to 0.007 ± 0.0033 in Drake Passage (Jullion et al., 2013). This is equivalent to 73 ± 26 Gty^{-1} of freshwater, and has been linked to increased Antarctic continental glacial melt, representing around half of existing estimates (Rignot et al., 2008). In some sectors AABW freshening may be accelerating (Menezes et al., 2017). *Medium confidence* is ascribed to the overall freshening trend, but sparse sampling means that assertions of its acceleration have *low confidence* as higher temporal resolution observations observe significant interannual variability in AABW properties at other export locations (Meijers et al., 2016).

3.3.1.2.3 Stratification

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

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

For the Southern Ocean, there is only limited information concerning stratification changes in the post-AR5 period. An increase in stratification caused by strengthened discharge of freshwater from the Antarctic Ice Sheet was invoked as a mechanism to suppress vertical heat flux and permit an increase in sea ice extent (Bintanja et al., 2013), though most studies conclude that glacial freshwater input is insufficient to cause a significant expansion (Swart and Fyfe, 2013; Pauling et al., 2017) (see also Section 3.3.1.1). Schmidtko et al. (2014) noted a moderate shoaling of the Winter Water layer (the summer subsurface temperature minimum layer in the Southern Ocean) over the period 1975-2012, concentrated in the Indian and Southeast Pacific sectors.

3.3.1.2.4 Biogeochemistry, carbon and ocean acidification

Since AR5, new observations have demonstrated the spatial and temporal variability of ocean acidification and controlling mechanisms of carbon systems in different regions. In the Canada Basin, 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 degrees, and deepened by about 100 m between the 1990s and 2010. In the East Siberian Arctic Shelf, extreme aragonite undersaturation was observed, reflecting pH changes in excess of those projected in this region for 2100 (Semiletov et al., 2016), and this feature was also observed along the continental margin and traced in the deep Makarov and Canada Basins (Anderson et al., 2017a). Persistent acidification here is driven by the degradation of terrestrial organic matter and discharge of Arctic river water with elevated CO_2 concentrations (*high confidence*), rather than by uptake of atmospheric CO_2.

The dissolved inorganic carbon (DIC) concentration increased in the subsurface waters (150-1400m) in the central Arctic between 1991 and 2011 (*high confidence*) (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 μmolkg^{-1} yr^{-1}in the upper Polar Deep Water due to anthropogenic CO_2, while no trend was observed in nutrient concentrations in the same water masses. In waters deeper than 2000 m, no significant trend was observed for DIC and nutrient concentrations. Observation-based estimates 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)* (MacGilchrist et al., 2014).

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

The major advance in understanding of CO_2 fluxes in the Southern Ocean since AR5 is from the decadal mean estimate (~$1PgCy^{-1}$ ± 0.5) and linear response to increasing anthropogenic CO_2 prior to 2013 (Takahashi et al., 2012; Lenton et al., 2013) towards new constraints of its seasonal-to-decadal variability (McNeil and Matear, 2013; Landschützer et al., 2014; Landschützer et al., 2015; Ritter et al., 2017; Gregor et al., 2017a; Gregor et al., 2017b). This advance has provided new insight to the earlier model-based assessment of a weakening CO_2 sink in the 1990s (Le Quéré et al., 2007), revealing that it was part of a decadal cycle that reversed in the 2000s (Landschützer et al., 2015; Munro et al., 2015; Williams et al., 2017). Resolving the decadal modes of variability has shown that the mean annual flux anomaly of CO_2 in the Southern Ocean can vary from approximately $0.3PgCy^{-1}$ ± 0.1 in 2001-2002 to $-0.4PgCy^{-1}$ in 2012 (Landschützer et al., 2015). The decadal mode appears to be linked to interannual adjustments in winter maxima possibly linked to the SAM (Landschützer et al., 2015; Gregor et al., 2017a) whilst summer ingassing variability may be linked to adjustments in primary productivity associated with ENSO (Conrad and Lovenduski, 2015) (Section 3.A.3.2). The decadal structure has the potential make a significant contribution to the magnitude and timing of the "missing" carbon determined in the global carbon budget (Le Quéré et al., 2017).

An additional driver that has emerged from increasing anthropogenic CO_2 fluxes is changes to the buffering capacity of the Southern Ocean; this has started to increase the amplitude of the seasonal cycle of pCO_2 over the past 3 decades (1.1 ± 0.3 $\mu atm.dec^{-1}$) (McNeil and Sasse, 2016; Landschützer et al., 2018) (Section 3.A.3.3). The confidence levels for the decadal modes and the trends in decreasing buffering capacity are *medium* to *high,* but data sparseness and model limitations make the confidence on potentially important links to seasonal drivers *low* to *medium*.

Recent reassessments of carbon storage in the Southern Ocean reveal strong sensitivity to changes in meridional overturning circulation, with anthropogenic and natural carbon being highly variable (± 50%) but out of phase on decadal timescales (DeVries et al., 2017; Tanhua et al., 2017); see also Section 3.A.3.4). Both mode and intermediate waters (SAMW and AAIW) are especially influential in this changing storage, also showing a high sensitivity to meridional shift in the wind stress (Swart et al., 2015; Tanhua et al., 2017). Zonal structure in the variable uptake and storage of anthropogenic carbon is not well resolved; the presence of subduction hotspots that suggest that basin-wide studies may be underestimating the importance of SAMW subduction as a principal storage mechanism has been highlighted (Langlais et al., 2017). The confidence levels on the decadal variability and wind sensitivity of natural and anthropogenic carbon in the Southern Ocean are *medium* to *high*.

Contemporary variability and trends in ocean carbonate chemistry that are consistent with Southern Ocean acidification have been observed (Mattsdotter Björk et al., 2014; Freeman and Lovenduski, 2015; Munro et al., 2015) and modelled (Sasse et al., 2015; McNeil and Sasse, 2016). Current estimates of the strengthening impacts of Southern Ocean acidification are best illustrated by the 3.9 ± 1.3% decrease in derived calcification rates (1998 – 2014) (Freeman and Lovenduski, 2015). These changes have strong regional character with decreases in the Indian and Pacific Sectors (7.5-11.6%) and increases in the Atlantic Ocean (14.3 ± 5.1%). This period coincides with the invigoration of CO_2 uptake by the Southern Ocean (Landschützer et al., 2015; Gregor et al., 2017a) but its regional character highlights that long-term trends are a complex interplay of regional ecological, biogeochemical and physical drivers.

A particular aspect of polar marine biogeochemistry is the potential for significant nutrient and organic carbon delivery from cryospheric sources, including subglacial meltwater, icebergs, surface runoff and melting of the base of ice shelves (Wadham et al., 2013; Hood et al., 2015; Raiswell et al., 2016; Hodson et al., 2017) (Figure 3.9). There is *medium evidence* that marine-terminating glaciers indirectly amplify nutrient fluxes by stimulating upwelling of nutrient-replete ocean water at the calving front (Meire et al., 2017a) and because of high carbon/nutrient burial and recycling rates in fjords (Wehrmann et al., 2013; Smith et al., 2015a). There is *high agreement* based upon *medium evidence* that changes in nutrient and organic matter export from ice sheets will impact wider biogeochemical cycles and ecosystem services (e.g. fisheries) (Hood et al., 2015; Milner et al., 2017). However, there is *limited evidence* for the scale and geographical distribution of these impacts (Meire et al., 2017a; Milner et al., 2017). *Limited evidence* indicates dissolved nutrient fluxes from the Greenland Ice Sheet increase during high melt years, but that the response of the dominant sediment-bound fraction is complex and may not increase with rising melt (Hawkings et al., 2015). Thus, there is *low confidence* overall in the magnitude of the response of nutrient fluxes from ice sheets to enhanced melting. A confounding influence to assessments is the landward march of marine-terminating glaciers and collapse of ice shelves, which is already observed in some climatically sensitive regions (Cook et al., 2016). This has the potential to drive major shifts in nutrient supply to coastal waters (Figure 3.9). *Limited evidence* suggests that heightened erosion of unconsolidated sediments in expanding proglacial zones (Monien et al., 2017) and increased diffuse nutrient fluxes from newly exposed glacial sediments on the seafloor (Wehrmann et al., 2014) (Figure 3.8) will amplify nutrient supply, whilst other nutrient sources may be cut off (e.g., icebergs, upwelling of marine water; Meire et al. (2017a)). There is *high agreement* based upon *limited evidence* that this will alter food supply to higher trophic levels of marine food webs (Meire et al., 2017a; Milner et al., 2017) (see also Section 3.3.3).

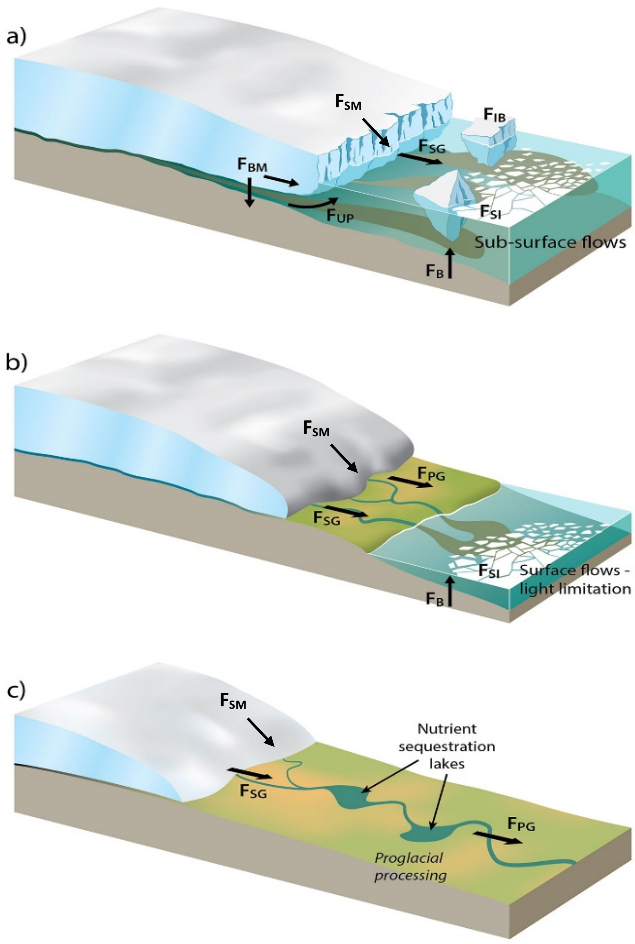


Figure 3.9: Potential shifts in nutrient fluxes (F) with landward retreat of marine-terminating glaciers (a) at different stages (b and c) (BM= basal melting, SG/SM=subglacial/surface melt, IB =icebergs, SI=sea ice, B=benthic (sea-floor), PG=proglacial).

3.3.1.3 Ocean Circulation

3.3.1.3.1 General circulation

Satellite radar altimetry information indicates a general strengthening of the surface geostrophic currents in the Arctic basin. Between 2003 and 2014, the strength of the Beaufort Gyre circulation approximately doubled, with similar increases in the strength of the southward surface flow at Fram Strait. In both regions, current speeds increased from around 2-4 cm/s to around 6-8 cm/s (*medium confidence*) (Armitage et al., 2017). Over 2001-2014, annual Bering Strait volume transport from the Pacific to the Arctic Ocean increased from 0.7×10^6 m^3s^{-1} to 1.2×10^6 m^3s^{-1} (*medium confidence*) (Woodgate et al., 2015).

Whilst the relative invariance of the ACC transport appears to have persisted post-AR5 (*medium confidence*) (Chidichimo et al., 2014; Donohue et al., 2016), new insights into the variability of transport of the subpolar regions of the Southern Ocean has been obtained. Using altimetry for the period 2011-2016, Armitage et al. (2018) observed significant responses of the Weddell and Ross Gyre transports to changing wind forcing (*high confidence*), with both SAM and ENSO implicated.

3.3.1.3.2 Eddy variability

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

Since AR5, there is increased evidence that the mesoscale eddy field in the Southern Ocean is intensifying, probably in response to wind energy input (*medium confidence)*. Hogg et al. (2015) used satellite altimetry data to demonstrate an increase in eddy kinetic energy in the Pacific and Indian sectors of the ACC of 14.9 ± 4.1 cm^2s^{-2} per decade and 18.3 ± 5.1 cm^2s^{-2} per decade respectively since the early 1990s. This is supported in eddy resolving models, which also show such a relationship has marked regional variability (Patara et al., 2016).

3.3.1.3.3 Overturning circulation and water mass production

Arctic processes, such as the discharge to the ocean of freshwater from the Greenland Ice Sheet, have the potential to impact on the formation of the headwaters of the Atlantic Meridional Overturning Circulation (AMOC; see Chapter 6). In parallel, Southern Ocean overturning circulation is the mechanism by which much of the global deep ocean is renewed. It is challenging to measure the Southern Ocean overturning directly, and the upper cell was incorrectly reported in AR5 as having slowed. However, indirect estimates since AR5 provide support for the increase in the upper ocean overturning proposed by Waugh et al. (2013). Waugh (2014) and Ting and Holzer (2017) suggest that over the 1990s-2000 water mass ages changed in a manner consistent with an increase in upwelling and overturning. However, inverse analyses suggest that overturning experiences significant inter-decadal variability in response to wind forcing (DeVries et al., 2017); combined with the indirect nature of observations, there is *low to medium confidence* in there having been an acceleration in overturning.

The lower limb of the Southern Ocean overturning circulation is primarily associated with AABW production and export from the Antarctic margins; available evidence indicates that the volume of this water mass has decreased *(medium confidence)* (Purkey and Johnson, 2013; Desbruyeres et al., 2017), thinning at a rate of 8.1 m yr^{-1} since the 1950s (Azaneu et al., 2013). This suggests that AABW export has likely slowed, though direct observational is difficult to obtain (*low confidence)*. The large-scale impacts of AABW changes include a potential influence on the strength of the AMOC (e.g. Patara and Böning (2014); also Chapter 5).

3.3.1.3.4 Movements of fronts and current cores

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. Since then, substantial contrary evidence has emerged. Gille (2014) computed displacements of a transport-weighted index of mean ACC position, and found no long-term trend and no statistically significant correlations with winds. Similar results were obtained via skewness- (Shao et al., 2015), wavelet- (Chapman, 2017) and kinetic energy- (Chambers, 2018) based studies. The discrepancy between these studies and those assessed in AR5 appears to be caused by issues associated with using a fixed sea surface height contour as a proxy for frontal position in the presence of strongly eddying fields (Chapman, 2014) and large-scale trends in sea surface height due to steric change. These recent findings do not preclude more local changes in frontal position, but the likelihood of there having been a net southward movement of the mean ACC is here reassessed as having *low confidence*.There is comparatively little knowledge on changing Arctic frontal positions and current cores since AR5, with the exception that the center of the Beaufort Gyre in 2013 was located about 300 km to the northwest of its position in 2003, contemporaneous with changes in its freshwater accumulation and alterations in wind forcing (Section 3.3.1.2.2) (Armitage et al., 2017).

3.3.2 Projected Changes in Ocean and Sea Ice

3.3.2.1 Model Projections of Sea Ice

Historical simulations from CMIP5 models capture observed declines in sea ice extent and thickness (Massonnet et al., 2012; Stroeve et al., 2012b; Stroeve et al., 2014b; Stroeve and Notz, 2015), though the sea ice thickness patterns, general features of Arctic atmospheric circulation, and ice drift rates are not well simulated (Stroeve et al., 2014b) (*high confidence*). Arctic sea ice extent loss scales linearly with both global temperatures and cumulative CO_2 emissions in both simulations and observations, suggesting that climate models realistically capture the integrated climate sensitivity of sea ice to climate change. However, the modeled sensitivity (ice loss per unit of warming) is about half that of observations, a result of the models underestimating the increase in downwelling longwave radiation associated with increases in atmospheric CO_2 (Notz and Stroeve, 2016b).

CMIP5 models project continued declines in Arctic sea ice through the end of the century (Overland and Wang, 2013; Notz and Stroeve, 2016b). There is a large spread in the timing of when the Arctic may become ice free in the future (Massonnet et al., 2012; Stroeve et al., 2012b) as a result of internal climate variability (Notz, 2015; Swart et al., 2015), scenario uncertainty (Stroeve et al., 2012b; Liu et al., 2013), and model uncertainties related to sea ice dynamics (Rampal et al., 2011). Internal climate variability alone results in an an uncertainty of 21 years in the timing of seasonally ice-free conditions (Jahn et al., 2016). The clear link between the summer sea ice extent and cumulative CO_2 emissions provide a basis for when ice-free conditions may be expected. With an additional 1000 Gt of CO_2, the Arctic is *very likely* to become ice-free in September (Notz and Stroeve, 2016b). At current emission rates of 35 to 40 Gt of CO_2 yr^{-1}, this will happen before the middle of the century, in agreement with a 2°C global temperature increase (Mahlstein and Knutti, 2012). On the other hand, for emissions compatible with a 1.5°C global warming target, sea ice in September is *very likely* to survive (Notz and Stroeve, 2016a; Jahn, 2018; Sigmond et al., 2018).

CMIP5 models show a wide range of mean states and trends in Antarctic sea ice (Turner et al., 2012; Shu et al., 2015). Ensemble means across multiple models show a decrease in total Antarctic sea ice extent during the satellite era, in contrast to the observed increase. This difference can be explained by internal variability (Polvani and Smith, 2013; Zunz et al., 2013), however, interannual sea-ice variability in the models is much larger than observed (Zunz et al., 2013), which may mask actual disparity between models and observations. 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 understanding of these issues, in part because there is no consensus on the drivers of observed changes (see Section 3.3.1.4.1), and also because there are a number of model biases that may explain sea ice error. Climate models tend to be too stratified in the Southern Ocean (Sallée et al., 2013b) which would diminish the delayed surface response, making sea ice too responsive to greenhouse gas forcing. Inadequate representation of cloud processes means that models have a warm surface bias in the Southern Ocean (Schneider and Reusch, 2015b). Models tend to underestimate meridional wind variability, limiting the sea ice response in the regions with the greatest observed trends (Purich et al., 2016b; Schroeter et al., 2017). Westerly wind changes are also underestimated by the models (Purich et al., 2016a).

IPCC AR5 reported that Antarctic sea ice extent is projected to decline over the 21st century, with the magnitude of decrease dependent on emissions scenario (i.e., greater decline under RCP8.5 than RCP4.5). Since AR5, research has focussed on explaining observed and simulated trends over the historical period, with little new research on projections. In addition to the issues outlined above, freshening by melt water from ice shelves could mitigate future sea ice loss (Bintanja et al., 2015), a process which is not represented in CMIP5 models. Due to the known biases and disagreement with observed trends in the CMIP5 models, there is *low confidence* in our ability to make reliable projections of Antarctic sea ice. This uncertainty reduces confidence in projections of Antarctic Ice Sheet surface mass balance, because sea ice biases affect Antarctic temperature and precipitation trends (Bracegirdle et al., 2015). They may also impact projected changes in the Southern Hemisphere atmosphere jet (Bracegirdle et al., in press), with implications for the Southern Ocean overturning circulation and the Antarctic Circumpolar Current.

[START BOX 3.2 HERE]

Box 3.2: Polynyas

Arctic Coastal Polynyas

Polynyas (areas of open water surrounded by sea ice) form regularly in many Arctic regions during winter and spring due to a combination of latent (wind) and sensible (heat) effects (Barber and Massom, 2007), and are areas of intense air-ice-ocean exchange (Morales Maqueda et al., 2004). The warm and exposed ocean surface creates very high heat fluxes and sea ice formation rates during winter, releasing brine and creating dense water that helps ventilate the stratified Arctic Ocean (Barber et al., 2012). Polynyas are projected to change in different ways depending on regional ice conditions and the processes responsible for formation. They may cease to exist where seasonal sea ice disappears, or evolve to become part of a marginal sea ice zone due to changes in ice dynamics (i.e., the North Water polynya and the Circumpolar Flaw Lead). Further reductions in sea ice are projected for Arctic shelf seas which have lost ice in recent decades (Onarheim et al., 2018). By 2100 under RCP8.5, all of Alaska's northern shore is projected to be ice-free all year, as are the Kara and Barents Seas and Baffin Bay, while the Siberian coast still has approximately six months of sea ice cover (Barnhart et al., 2015). New or enlarged polynyas could result in regions where thinner ice becomes more effectively advected offshore, or where marine terminating glaciers increase land ice fluxes to the marine system.

In spring and summer, polynyas are the first areas exposed to solar insolation. The spring phytoplankton bloom therefore starts earlier here, and the ocean is well-ventilated and often nutrient rich, so the entire biological range from phytoplankton to marine mammals thrive in polynya waters. Early and sustained phytoplankton blooms are a key feature of polynyas so long as nutrients and light are available to the euphotic zone. Secondary production and upper food web processes usually have adapted to the early availability of energy to the system with arrival of higher trophic species having adapted to the early availability of energy (Asselin et al., 2011).

Because of the abundant availability of marine resources including seals, whales and fish in and around polynyas, they have been regular areas for hunting by Arctic peoples over thousands of years (Barber and Massom, 2007). Recent implementation of Inuit led marine management areas acknowledge the Inuit traditional knowledge of polynyas and recognize that development of fisheries and non-renewable resources such as oil and gas are possible in polynya systems. The Inuit Circumpolar Council's Pikialasorsuaq Commission is an example of a proposal to develop an Inuit management area in the North Water Polynya (see Cross Chapter Box 3).

Antarctic Coastal Polynyas

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

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

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

The productivity associated with these polynyas is a critical food source for some of the most abundant top predators in Antarctic waters, including penguins, albatross, and seals (Raymond et al., 2014; Labrousse et al., 2017; Malpress et al., 2017). However, only a fraction of the carbon fixed by phytoplankton in coastal polynyas is consumed by upper trophic levels. The rest either sinks to the seafloor where it is remineralized or sequestered (Shadwick et al., 2017), or is advected off the shelf (Lee et al., 2017b). Given the high amount of residual macronutrients in polynya surface waters, future changes in ice shelf melt rates could increase water column productivity 1.7-fold (Alderkamp et al., 2015), dramatically influencing Antarctic coastal ecosystems and the ability of continental shelf waters to sequester atmospheric carbon dioxide (Arrigo and van Dijken, 2015).

The Weddell Polynya

The Weddell Polynya is a large area of open water within the winter ice pack of the Weddell Sea (at approximately 60^{o}S, 15^{o}W). The polynya opens intermittently, and remained open for several years between 1974-1976, with an area of 0.2-0.3 million km^2 (Carsey, 1980). An area of low sea ice concentration has appeared in this area, following the extreme low Antarctic sea ice extent in spring 2016, but it remains an open question whether a persistent polynya will form again.

The polynya has so far formed close to the Maud Rise seamount, and may be caused by ocean eddies creating sea ice divergence over deep ocean water (Holland, 2001). This is unusual in Antarctica, where most polynyas form along the coast and are wind-driven, which result in overall ice production rather than a net melt of sea ice. Around Maud Rise, the ocean is weakly stratified, and sea ice formation causes mixing of warm, deep waters to the surface, sufficient to melt newly-formed sea ice (Martinson et al., 1981). This process may allow the Weddell Polynya to persist for some years, and causes 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).

CMIP5 models suggest that Weddell polynyas are a very common feature, with approximately half of the models hosting persistent polynyas during the historical period (Sallée et al., 2013b). In some models, phases of polynya activity appear for decades or centuries at a time, and then cease for a similar time (Reintges et al., 2017b). 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 model stratification due to lack of realistic freshwater input from ice shelves and melting ice bergs, likely offsetting sea ice volume and producing a *low confidence* in the future Weddell Polynya projections (Reintges et al., 2017a).

[END BOX 3.2 HERE]

3.3.2.2 Ocean Properties and Circulation

Consistent with the projected sea ice decline, there is *high confidence* that the Arctic Ocean will warm significantly towards the end of this century at the surface and in the deeper layers. Most CMIP5 models are able to capture the seasonal changes in surface heat and freshwater fluxes for the present day climate, and show that the excess summer solar heating is stored in the form of melting sea ice rather than increased ocean temperature (Ding et al., 2016). There is however large model bias for the present climate and little consensus among the models concerning whether warming will occur by a reduction of the heat loss at the surface, or by an increased ocean heat transport. Using RCP8.5, Vavrus et al. (2012) found that the Atlantic layer temperature will 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 with a lower Atlantic Water layer warming was found by Koenigk and Brodeau (2014) for RCP2.5 (+0.5°C), RCP4.5 (+1.0°C) and RCP8.5 (+2.0°C).

Comparing 26 different CMIP5 models for RCP4.5, Burgard and Notz (2017) found that changes in the ocean heat transport explain the Arctic Ocean warming for the multi-model mean, but that this is not a consensus across the models. Differences in ocean heat transport between models are largely compensated by changes in the net atmospheric surface flux, and most models show evidence of future Bjerknes compensation (Bjerknes, 1964) – with lower atmospheric heat transport towards the Arctic when the ocean heat transport is large, and vice versa. Comparing 20 CMIP5 models for RCP8.5, Nummelin et al. (2017) found a range in Arctic amplification between 2° and 6° north of 70°N consistent with, and associated with, increased ocean heat transport towards the Arctic.

A consistent but uncertain mechanism of the Arctic Ocean warming over the twenty first century is an increased ocean heat transport into the Barents Sea (Koenigk and Brodeau, 2014). Onwards from about 2050, this becomes ice-free during winter (Onarheim and Årthun, 2017), and the main response will be an increased ocean to atmosphere heat flux and related surface warming (Smedsrud et al., 2013). When the winter sea ice disappears the heat loss cannot increase further, and the excess ocean heat will then contribute to the warming of the Atlantic Water layer inside the Arctic Basin (Koenigk and Brodeau, 2014). The ocean heat transport will also increase through the other Arctic gateways (Bering Strait, Fram Strait, and the Canadian Archipelago), but the increase appears smaller than in the Barents Sea.

The Arctic Ocean is expected to freshen at the surface in the future due to increased river runoff because of an intensified hydrological cycle (*medium confidence*)(Haine et al., 2015a). The related increase in stratification will likely contribute to the warming of the Atlantic Water layer at depth as upward vertical mixing will be reduced (Nummelin et al., 2016b). There are however systematic salinity biases (~1 psu) in the CMIP5 models for the present day climate, with all models being too saline at the surface in the Canada Basin, and too fresh at depths between 50-400 m all across the Arctic Basin (Ilicak et al., 2016).

CMIP5 modelling indicates that observed Southern Ocean warming trends will continue under RCP4.5 and RCP8.5, leading to 1°C–3°C warming by 2100 mostly in the upper ocean (Sallée et al., 2013b). Model projections demonstrate a similar distribution of heat storage to historical observations, notably focused in deep pools north of the Subantarctic Front (e.g., Armour et al., 2016). AABW becomes coherently warmer by up to 0.3°C by 2100 across the model ensemble under RCP8.5 (Heuzé et al., 2015). The upper ocean water masses also become considerably fresher (0.1 psu) (Sallée et al., 2013b) with an overall increase in stratification and shoaling mixed layer depths (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 biases in historical water mass properties (Sallée et al., 2013b) and sea surface temperature, which may be up to 3°C too warm in the historical runs (Wang et al., 2014).

Significant uncertainties in Southern Ocean / global climate interactions remain. Eddy parameterisations and background stratification significantly affect ocean heat uptake efficiency (Downes et al., 2018), explaining up to 50% of global ocean heat uptake variability (Kuhlbrodt and Gregory, 2012) and there is a complex dynamical relationship between heat uptake and subsequent modification of storage by changed ocean circulation (Cheng et al., 2013). Ocean-atmosphere feedbacks are also poorly understood; Frölicher et al. (2015) find little climate feedback between the ocean and atmosphere on centennial timescale despite a wide range of modelled ocean heat uptakes. On decadal timescales however, extra heat uptake by the ocean has been suggested as a possible driver for the so-called 'hiatuses' in surface atmosphere warming over the last century. Llovel et al. (2014) find a net ocean warming equivalent to a radiative imbalance of 0.64 ± 0.44 W m^{-2} since 2005, which balances the top-of-atmosphere radiative imbalance of 0.50 ± 0.43 W m^{-2} for the period from 2001 through 2010 (Loeb et al., 2012). The dominance of Southern Ocean heat uptake means that changes in deep Southern Ocean heat content are believed to be significant contributors to such interdecadal variability (Chen and Tung, 2014).

The considerable CMIP5 intermodel variability in Southern Ocean circulation projections reported in AR5 (Meijers et al., 2012; Downes and Hogg, 2013) is largely unchanged. While the northern boundary of the ACC is predicted to shift polewards in these models, the ACC core does not demonstrate any coherent change (*moderate confidence*). There are significant differences in subpolar gyre strength and position response also, however these vary greatly by region and model with no coherent ensemble signal (Meijers et al., 2012) and there is *low confidence* in projected ACC transport or subpolar gyre circulation changes.

The vertical circulation of the Southern Ocean is predicted to change in response to the positive SAM mode projected in climate warming scenarios. This may increase the upper cell subduction and northward transport by up to 20% (Downes and Hogg, 2013) but model performance is limited by representation of eddy processes (Gent, 2016; Downes et al., 2018). The formation and export of deep bottom waters is predicted to continue to decrease (Sallée et al., 2013b; Heuzé et al., 2015) due to warming and freshening of surface source waters near the continent. These are, however, some of the most poorly-represented processes in the simulated global ocean; *low confidence* is thereore ascribed to future Southern Ocean circulation and water mass projections.

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

CMIP5 models project that the uptake of CO_2 by the Southern Ocean will increase from the contemporary 0.91$PgCy^{-1}$ to 2.38 (1.65-2.55) $PgCy^{-1}$ by 2100, but the growth in uptake will stop in about 2070 corresponding to cumulative CO_2 emissions of 1600GtC (Kessler and Tjiputra, 2016; Wang et al., 2016b). The onset of aragonite undersaturation in the Southern Ocean is influenced by the seasonal cycle of carbonate as well as the influence of anthropogenically-forced reduced buffering trend on the seasonal cycle (Sasse et al., 2015; McNeil and Sasse, 2016). The importance of the seasonal cycle is apparent when considering the year of onset of month-long and annual-mean undersaturation for the Southern Ocean under different scenarios: a sharp tipping point exists between RCP2.6 and RCP4.5/RCP8.5, with the latter two scenarios leading to the onset of pervasive mean annual undersaturation within 10 to 20 years of the onset of monthly undersaturation (Appendix 3.A, Table 2). RCP2.6 results in a 99.8% reduction of the area impacted by seasonal undersaturation (Appendix 3.A, Figure 6) (Sasse et al., 2015). The existence of the tipping point is supported by predictions based on RCP8.5, that because of reduced buffering capacity, the onset of month-long hypercapnia (pCO2 > 1000μatm) in the Southern Ocean will occur around 2080, and that by 2100 almost the whole Southern Ocean will be impacted (McNeil and Sasse, 2016). This implies that under RCP4.5/8.5 scenarios, not only will calcification be impacted but possible organism physiology across ecosystems (Sasse et al., 2015). Despite the importance of the seasonal cycle, recent studies highlight that the importance of interannual variability driven by large scale atmospheric modes (ENSO and SAM) should be included in the predictions for the onset of both undersaturation and hypercapnia (Conrad and Lovenduski, 2015). Although the confidence level for the onset of reduced buffering capacity and undersaturation is *high* to *very high*, the model projections are still temporally and spatially uncertain so the overall confidence levels are *medium* to *high*.

One of the most important and likely additional outcomes from decreasing buffering capacity of the ocean is an amplification of the seasonal variability of pCO2 and pH (Hauck and Volker, 2015; McNeil and Sasse, 2016; Landschützer et al., 2018). The amplification accelerates the onset of hypercapnia (pCO_2> 1000uatm) to nearly 2 decades ahead of atmospheric forcing (McNeil and Sasse, 2016). Under RCP8.5, the Southern Ocean will be exposed to the dual effects of undersaturation and hypercapnia (Hauck and Volker, 2015; Sasse et al., 2015; McNeil and Sasse, 2016).

3.3.3 Implications for Marine Ecosystems

Climate change impacts on the polar ocean and cryosphere can have profound implications for marine ecosystems, with consequences for production and distribution at different trophic levels both in the pelagic and benthic realm (*high confidence*). The impacts of climate change on polar marine ecosystems would be spatially heterogeneous with respect to the rate and severity of change (*high confidence)*.

3.3.3.1 Arctic

We follow the definition of the marine Arctic as given by PAME (2013), which comprises the areas of the Arctic Large Marine Ecosystems [Figure 3.x]. The large ecoregions, identified by Carmack et al. (2015) provide a framework for our discussion of regional heterogeneity in impacts (Carmack et al., 2015; Wassmann et al., 2015).

3.3.3.1.1 Arctic lower trophic level responses

Primary production

There is evidence that the combination of loss of sea ice, freshening, and regional stratification (3.3.1.) has affected the timing, distribution and production of lower tropic level species (*high confidence*). Data from Earth observing satellites 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 a shift in seasonal phytoplankton bloom phenology, with spring blooms coming earlier (Kahru et al., 2011) and being dominated by larger-celled phytoplankton (Fujiwara et al., 2016). The longer open water season in the Arctic has also increased the incidence of fall blooms, a phenomenon rarely observed in Arctic waters previously (Ardyna et al., 2017).

Perhaps the most dramatic consequence of thinner Arctic sea ice cover has been the intense phytoplankton blooms that develop beneath first year sea ice. Observed in detail for the first time in the Arctic in 2011 (Arrigo et al., 2012), blooms of this size and intensity were thought to be restricted to the marginal ice zone and the open ocean where ample light reaches the surface ocean for rapid phytoplankton growth. We now know that these blooms can thrive beneath the sea ice in areas of reduced sea ice thickness, increased coverage of melt ponds (Arrigo et al., 2012; Arrigo et al., 2014; Zhang et al., 2015; Jin et al., 2016; Horvat et al., 2017) and large lead fractions (Assmy et al., 2017), although the latter has not changed significantly in the last three decades (Wang et al., 2016a).

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

In addition to its impact on phytoplankton bloom dynamics, the decline in the proportion of multiyear sea ice and proliferation of a thinner first year sea ice cover may favor growth of microalgae within the ice due to increased light availability. 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).

All of these ongoing phenological changes in NPP are impacting the biogeochemistry and ecology of the Arctic Ocean. 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).

Zooplankton

The phenology, magnitude and duration of zooplankton production and the zooplankton community composition in the Arctic are changing in response to increased water temperatures (3.3.1) and spatial pattern and timing of the ice algal and phytoplankton blooms (*medium confidence*). At the more southern boundaries of the Arctic such as the Bering Sea, warming conditions have led to a reduced production of large copepods and euphausiids, with consequences to fisheries (Sigler et al., 2017; Kimmel et al., 2018). On more northern shelves, the increased open water period may have led to long-term increases in large copepods within the Chukchi (Ershova et al., 2015) and Beaufort Seas (Smoot and Hopcroft, 2017), and in the Central Basins zooplankton biomass has increased (Hunt et al., 2014).

Recent studies suggest that the Chukchi Sea may be transitioning from the benthic-dominated system of the past, to a more pelagic-dominated one (Moore and Stabeno, 2015). Projections based on the SRES scenario A1B, suggest that large changes in the production, distribution and magnitude of the keystone copepods *Calanus finmarchicus* and especially *C. glacialis* in the Eurasian Arctic will occur towards the end of the century (Wassmann, 2015). Other models have also suggested *C. glacialis* has, and should continue to, benefit from a warmer Arctic Ocean (Feng et al., 2018), while in the transition zone between Arctic and Atlantic water masses, they may face increasing competition from *C. finmarchicus* (Dalpadado et al., 2016). The same study of euphausiid (krill) and amphipod dynamics in Kongsfjorden, Spitsbergen (79°N) and adjacent waters, revealed that if projected warming trends persist the Atlantic/boreal euphausiid species will be favoured, while Arctic species, such as the amphipod *Thermisto libellula*, may decline (Dalpadado et al., 2016).

Seasonal and spatial heterogeneity in the presence of undersaturated waters is expected in Polar Regions with marked differences in projected extent under different RCPs (3.3.1.2) with associated impacts on calcifying zooplankton and pelagic mollusks (Larsen et al., 2014b; Howes et al., 2015). Recent studies provided evidence that pteropods have natural defense mechanisms that may allow repair of shells damaged by exposure to waters undersaturated with respect to aragonite (Peck et al., 2018). In contrast, ocean acidification is expected to negatively impact survival of some crab and shellfish species in the future, current ocean conditions do not appear to have negatively impacted crab production in the Bering or Barents Seas (Mathis et al., 2015; Punt et al., 2015).

Ecosystem effects of changes in glacial systems

There is *high agreement* based upon *medium evidence* that changes in nutrient and organic matter export from ice sheets will impact wider biogeochemical cycles and ecosystem services (e.g. fisheries) in some way (Hood et al., 2015; Milner et al., 2017). However, there is *limited evidence* for the scale and geographical distribution of these impacts (Meire et al., 2017a; 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) (*medium confidence*). There is *medium evidence* that marine-terminating glaciers indirectly amplify nutrient fluxes by stimulating upwelling of nutrient replete ocean water at the calving front (Meire et al., 2017a) and because of high carbon/nutrient burial and recycling rates in fjords (Wehrmann et al., 2013; Smith et al., 2015a). This process 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 the coastal zone around Greenland with potentially large ecological implications, negatively affecting production of commercially harvested fish (Meire et al., 2017b). Also, changing conditions in

Arctic fjords may suppress or expand the habitat and niche of planktonic organisms. A recent study by (Arrigo et al., 2017a) showed that melting glaciers could stimulate large summer phytoplankton blooms in southwest Greenland waters (*medium confidence*).

3.3.3.1.2 Arctic benthic communities

There is evidence that earlier spring sea ice retreat and later fall sea ice formation are changing the phenology of primary production with cascading effects on Arctic benthic community production (*medium confidence*). Benthic macrofauna (e.g., clams, worms, and amphipods) and more mobile megafaunal invertebrates (e.g., sea stars, crabs) are vulnerable to these changes because of their dependence on of high biological productivity (Link et al., 2013).

Over the last decade, a northward shift in the distribution of benthic species has been detected as well as subsequent changes in community composition, in the Arctic including the northern Bering Sea (Grebmeier et al., 2006; Grebmeier, 2012), Western Greenland (Renaud et al., 2015), and the Barents Sea (Jørgensen et al., 2012; Fossheim et al., 2015). Rapid and extensive structural changes in the rocky-bottom communities of two Arctic fjords in the Svalbard Archipelago have been documented during the period 1980-2010 and linked to gradually increasing seawater temperature and decreasing sea ice cover (Kortsch et al., 2012; Kortsch et al., 2015). Also, there is indication of declining benthic biomass in the northern Bering Sea (Grebmeier and Cooper, 2016) and southern Chukchi Sea (Grebmeier et al., 2015). However, biomass of kelps have increased considerably in the intertidal to shallow subtidal in Arctic regions over the last 2 decades, connected to reduced physical impact by ice-scouring and increased light availability as a consequence of warming and concomitant fast-ice retreat (see also Section 3.A.3.5) (Kortsch et al., 2012; Bartsch et al., 2016; Paar et al., 2016) (*medium confidence*).

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

3.3.3.1.3 Shifts in spatial distribution and production of Arctic fish

Recent observations support previous findings that a number of fish species in Arctic areas have changed their spatial distribution patterns substantially over the recent decades (*high confidence*). This includes ecological and commercially important stocks in the Bering and Barents Seas, while data is severely limited in most Arctic Ocean shelf regions (Box 3.3).

In the recent decade, there is evidence that climate variability has impacted the productivity of several commercially important marine fish in the Barents and Bering Seas with warm conditions favouring fish production in the Barents Sea (h*igh confidence*) whereas, warm conditions were associated with reduced production of gadids in the Bering Sea (*medium confidence*). Retrospective studies and laboratory experiments suggest that high lipid content zooplankton may be less abundant in warm ocean conditions in the Bering Sea resulting in reduced overwintering success of some arctic and sub-arctic species (Heintz et al., 2013). Time sereies on responses of anadromous fish in the high Arctic is limited, although these stocks are will also be exposed to a wide range of future stressors (Reist et al., 2016).

Evidence continues to support previous findings that interannual and decadal variability impacted the productivity (growth and reproductive success) of some marine fish in the Barents and Bering Seas (*high confidence*). The annual production of fish stocks in high latitudes is governed by a gauntlet of complex processes that impact stocks differently throughout the first year of life; many of these processes are influenced by temperature variability (Ottersen et al., 2014; Szuwalski et al., 2014). In the Barents Sea, heightened temperatures have expanded suitable feeding areas which also contributed to increased Atlantic cod (*Gadus moruha*) production (Kjesbu et al., 2014). In contrast, polar cod (*Boreogadus saida*) are expected to be negatively affected by a shortened ice-covered season and reduced sea-ice extent through loss of spawning habitat and shelter, increased predatory pressure, and reduced prey availability (Christiansen, 2017), and impaired growth and reproductive success (Nahrgang et al., 2014). There is also evidence that environmental variability influences the production of anadromous species such as Arctic char (*Salvelinus alpinus*), brown trout (*Salmo trutta*), salmon through its influence on environmetal stressors governing growth and winter survival (e.g., the "critical size and critical period" hypothesis) (Jensen et al., 2017).

Projected reductions in summer sea ice, increased stratification in summer (Section 3.3.1.1), shifting currents (Section 3.3.2.3) and fronts (Section 3.3.1.3.4) and increased ocean temperatures (Section 3.3.1.2.1) and ocean acidification (Section 3.3.1.2.3) are all expected to impact the future distribution of several marine fish and invertebrates (*high confidence*). Winter ocean conditions in the high Arctic are expected to remain cold limiting the immigration of resident populations of sub-arctic species on the high Arctic shelves, however seasonal advection of pelagic prey may allow feeding invasions to occur (Wassmann et al., 2015). Many demersal fish and invertebrates populations are constrained by the continental shelves and consequently they may not expand their habitat poleward beyond the shelf break. Therefore, the further expansion of Northeast Atlantic haddock (*Melanogrammus aeglefinus*) or cod is, expected to be limited to an eastward expansion along the Siberian shelf (Landa et al., 2014). The pelagic capelin are capable of entering the Polar Ocean, but they may be restricted in winter by availability of suitable spawning areas and lack of antifreeze proteins (Hop and Gjøsæter, 2013; Christiansen, 2017).

Under some RCPs, climate change will impact the future productivity of several marine fish stocks in the Arctic (*high confidence*). Regional climate scenarios, derived from downscaled global climate scenarios, have been used to drive environmentally linked fish population models with temperature-specific growth and predation rates to project the impacts of climate change on the production of southeastern Bering Sea groundfish (Hermann et al., 2016; Holsman et al., 2016; Ianelli et al., 2016). These scenarios project future declines in in the abundance of walleye pollock (*Gadus chalcogrammus*), Pacific cod (*G. microcephalus*) and arrowtooth flounder (*Atheresthes stomias*). Based upon downscaled projections from GCMs and a spatially explicit Individual Based Model (IBM), Hedger et al. (2013) predicted increases in Atlantic salmon (*Salmo salar*) abundance, both in marine and freshwater stages in northern Norway (river Alta around 70°N).

3.3.3.1.4 Shifts in production and spatial distribution of Arctic marine mammals and seabirds

Changes in the physical environment in the Arctic caused by global warming are resulting in distributional, phenological, behavioral and physiological changes in Arctic marine mammal and seabird populations and in the broader biotic communities that they occupy (*high confidence*; Gilg et al. (2012); Post et al. (2013); Meier et al. (2014); Laidre et al. (2015)). The cause of these changes include direct responses to habitat degradation induced by loss of sea ice, as well as responses mediated by changes in Arctic food webs or alterations to ecological interactions (and changes in human activities see Section 3.3.4).

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

In some species, shifts in distribution in response to changes in suitable habitat are associated with increased mortality. Increased mortality rates of walrus 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 because of 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 breeding ice for ice seals (Bajzak et al., 2011; Øigård et al., 2013) is occurring with increases in pup mortality and stranding in light ice years (Johnston et al., 2012; Soulen et al., 2013; Stenson and Hammill, 2014).

Climate impacts that reduce the availability of prey resources (via abundance declines or distribution shifts) can negatively impact marine mammals and seabirds (Asselin et al., 2011; Øigård et al., 2014; Hamilton et al., 2016; Brown et al., 2017b; Choy et al., 2017). Evidence suggests that ringed seals (*Pusa hispida*) in the marginal ice zone north of Svalbard are finding less sympagic food (less ice-associated diving) and are diving longer and deeper, resting less and searching more broadly, indicating increased foraging effort is required now compared to a decade ago (Hamilton et al., 2015). In some regions, ice declines have been associated directly with declines in body condition, ovulation, pregnancy rates and pup production as well as increased stress levels ; hunters also report more observations of sick seals in low ice conditions (Ferguson et al., 2017). 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, *Odobenus rosmarus* and gray whales, *Eschrichitus robustus*) (Brower et al., 2017; Udevitz et al., 2017; Szpak et al., 2018). Some species such as black-legged kittiwakes (*Rissa tridactyla*) show evidence of diet switching with a shift to a more diverse diet during a period of Arctic Atlantification (Section 3.3.1) (Vihtakari et al., 2018)

Changes in the timing, distribution and thickness of sea ice described in Section 3.4.1 and snow have been linked to phenological shifts, or redistribution, of denning or survival of polar bears (*Ursus maritimus*) (Derocher et al., 2011; Olson et al., 2017; Escajeda et al., 2018). Less ice (and more open water) is also driving polar bears to travel over greater distances and swim more than previously both in offshore and in coastal areas, which can be dangerous for young cubs (Durner et al., 2017; Pilfold Nicholas et al., 2017; Rode et al., 2018). Cumulatively, changes in sea ice patterns are driving demographic changes in polar bears, including declines in some populations where sea ice reductions are notable (Lunn et al., 2016; McCall et al., 2016). However, some polar bear populations are stable or increasing (Voorhees et al., 2014), even with regional declines in sea ice, because protective management measures have been successful in allowing severely depleted populations to recover despite habitat degradation (Aars et al., 2017) or because new food sources are suddenly available to polar bears (Galicia et al., 2016; Stapleton et al., 2016).

Time series going back to the 1980s only exist for a few locations in the Arctic for top predators, in these locations, the observed trends and adaptive capacity show regional differences across species (Laidre et al., 2015). Interdecadal comparisons of the body condition and productivity of two ice obligate pinniped species, ringed seals and bearded seals (*Erignathus barbatus*), in the Bering Sea, Chukchi Sea between 1975-1984 and 2003-2012 suggest that ringed seals have not been impacted by changes in the environment to date, while results for bearded seals were less certain, although no relationships between bearded seal biological parameters and sea ice were statistically significant (Crawford et al., 2015).

Changes in the spatial distribution of polar bears and killer whales can have top-down effects on other marine mammal prey populations (Reinhart et al., 2013; Øigård et al., 2014; Breed et al., 2017; Smith et al., 2017). Killer whales are sighted more often in the Canadian Arctic in recent decades and are thought to be spending longer seasons in the North due to declines in sea ice (Higdon et al., 2012).

There are some examples of adaptive behavioral responses in marine mammals to the extreme changes that have taken place in their habitats. For example, ringed seals in Svalbard are using terrestrial haul-out sites during summer for the first time in observed history, in the company of harbor seals in some areas, following major declines in sea ice (Lydersen et al., 2017). Although this shows capacity for flexibility, such behaviour does not enhance or sustain reproduction.

3.3.3.1.5 Ecosystem dynamics

Projected impacts of climate change are expected to alter the timing and flow and chemical properties of the Arctic with associated implications for the species composition, production and ecosystem structure of the high arctic (*medium confidence*) (Moore et al., 2016; Frainer et al., 2017). These changes will modulate the Atlantic and Pacific Arctic gateways to the broad shelf regions of the Arctic inflow ecosystems (Mueter et al., 2017). The consequences of increasing temperatures and reduced ice coverage in, e.g., the Barents Sea extend beyond habitat expansion of single species. The niche partitioning between sub-arctic 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). Northwards expansion of areas of Atlantic water masses causing affects the ecosystem through several pathways. Euphausiids and amphipods are major food of for Arctic fishes, and changes in the prey composition may have an impact on the feeding dynamics of these fish species (Dalpadado et al., 2016; Hunt et al., 2016c). Further, large piscivorous and semipelagic boreal species (like Atlantic cod) are replacing small-bodied benthivorous Arctic species in the northern Barents Sea, changing biogeography and ecosystem functioning (Figure 3.10) (Fossheim et al., 2015; Frainer et al., 2017). The capacity of species to adapt to changing prey availability varies.

In the Barents Sea in the Atlantic Arctic, evidence suggests that factors directly related to climate change (sea-ice dynamics, ocean mixing, bottom-water temperature change, ocean acidification, river/glacier freshwater discharge) are impacting benthic species composition (Birchenough et al., 2015). In addition, other human-influenced activities, such as commercial bottom trawling and introduction of non-indigenous species, are also regarded as major drivers of observed and expected changes in benthic community structure (Johannesen et al., 2017).

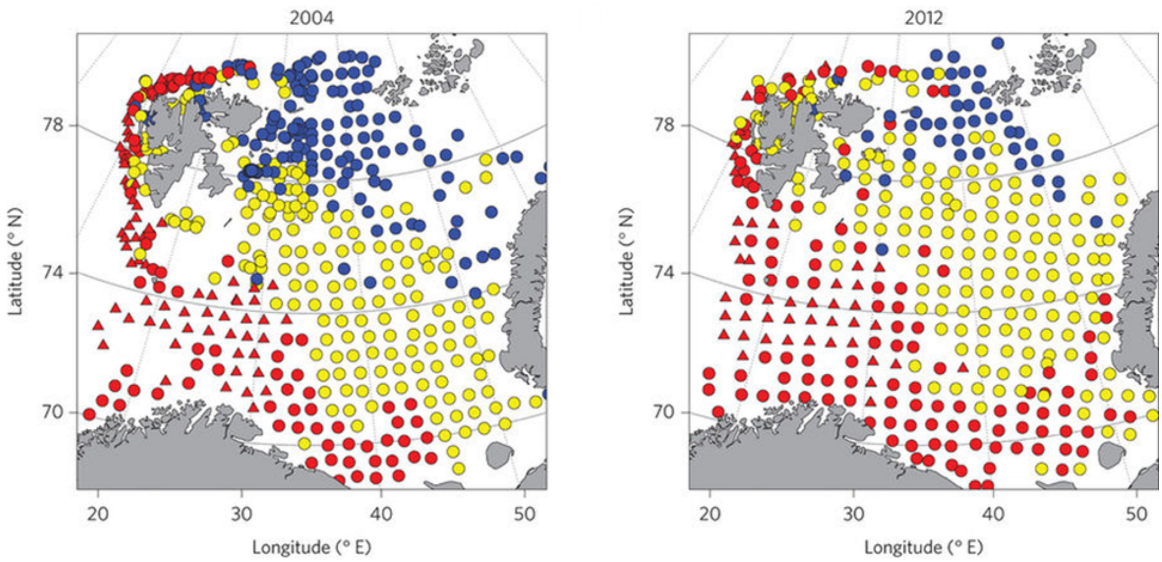


Figure 3.10: Changes in northern Barents Sea fish communities from 2004 (left panel) to 2012 (right panel). Observations from bottom trawl stations. Atlantic (red), Arctic (blue) and Central communities (yellow) symbols, respectively. Circles: shallow sub-communities, triangles: deep sub-communities. Adapted from Fossheim et al. (2015); Frainer et al. (2017).

3.3.3.2 Southern Ocean

As described in Chapter 28 of AR5 (Larsen et al., 2014a), marine ecosystem dynamics in the Antarctic region are dominated by the Antarctic Circumpolar Current (ACC) and its frontal systems, subpolar gyres, polar seasonality, the annual advance and retreat of sea ice, and the supply of limiting micronutrients for productivity (mostly commonly iron). Antarctic krill (*Euphausia superba*) play a central role in Southern Ocean foodwebs as consumers of productivity and as prey items for fish, squid, marine mammals and seabirds. This is in part due to their abundance and circumpolar distribution, although the abundance and, as evidence increasingly suggests, the importance of this species varies between different regions of the Southern Ocean (Constable et al., 2014; Larsen et al., 2014a; Siegel, 2016). New findings since AR5 characterise the nature of habitat change for Southern Ocean biota at the circumpolar scale (Gutt et al., 2015; Hunt et al., 2016a; Murphy et al., 2016; Gutt et al., 2017; Trebilco et al., In review), and the direct responses of biota to these changes (Table 3.1). These findings indicate that overlapping changes in key ocean and sea-ice habitat characteristics (temperature, sea-ice cover, ice-berg scour, mixed layer depth, aragonite under-saturation) will be important in determining future states of Southern Ocean ecosystems (Larsen et al., 2014a; Constable et al., 2016)(*medium confidence*). Indirect responses to physical change remain less well characterized because they are numerous and because it is challenging to determine the relative strength of positive and negative feedbacks which dictate the direction of indirect effects. Important advances have also been made in (i) identifying key variables to detect and attribute change in Southern Ocean ecosystem, as part of long-term (circumpolar) modelling designs (Cavanagh et al., 2017), and (ii) defining methods for using sea ice predictions from global climate models in ecological studies and in ecosystem models for the Southern Ocean (Larsen et al., 2014a).

Table 3.1: Summary of known direct responses of biota to changes in physical parameters in Antarctica and the Southern Ocean (based on Atkinson et al., 2004; Constable et al., 2017). UV = ultraviolet radiation. Acidification includes altered carbonate chemistry and pH. Sea-ice includes consideration of thickness, concentration, and extent without differentiating the factor/s causing change in each group of organisms. An upwards arrow indicates a positive relationship (increase in the physical variable is expected to cause an increase in the taxon). A downward arrow indicates a negative relationship (increase in the physical variable is expected to cause a decline in the taxon). A question mark (?) indicates where there is likely to be a response but the direction is uncertain, i.e. the result may be variable in space, time or for specific taxa, or the evidence is equivocal. As physical factors vary in their direction of change between different regions of the Southern Ocean, the responses in this table are used to interpret what specific directions of change may mean for the populations in a region. Indirect responses to physical parameters are addressed in the main text and are too numerous to capture in this table.

Taxon	UV	Temperature	Ocean acidification	Mixed Layer Depth	Sea-ice	Move with fronts	Eddies
Diatoms	↓	↑		↑	↓	↑	
Flagellates, *Phaeocystis*	↑	↓		↓	↓	↑	
Microzooplankton	?	↑	?		↓	↑	
Bacteria & viruses	↓	↑			↓	↑	
Zooplankton		↑				↑	
Salps					↓	↑	
Antarctic krill	↓	↓ subantarctic	↓		↑		
Nototheniid fish		↓					
Myctophid fish		↑				↑	
Oegopsid squid		↓ ?	↓ ?				
Southern Elephant seal					?		↑
Krill-eating seals				↓	↑ ?	↑	↑
King penguin						↑	↑
Emperor penguin					?		?
Adélie penguin				?	↑ no ice to lower ice conditions ↓ heavy ice conditions		
Macaroni penguin					↓		↑
Baleen Whales					?		?
Flying birds					↑ ?	↑ ?	↑ ?
Benthic communities		↓ ?	↓ ?				

3.3.3.2.1 *Southern Ocean primary production*

Changes in column-integrated phytoplankton biomass for the Southern Ocean (detected by space-based LIDAR) are coupled with changes in the spatial extent of ice-free waters, suggesting little overall change in biomass per area at the circumpolar scale (Behrenfeld et al., 2016) (*low confidence*). A similar trend of invariance between annual cycles of remotely sensed column-integrated primary production south of 50S was noted by Arrigo et al. (2008) for the period between 1998 and 2006, inclusive (*low confidence)*. At a regional scale, local-scale forcings (e.g. retreating glaciers and topographically steered circulation) and stratification are key determinants of phytoplankton bloom dynamics at coastal stations on the Western Antarctic Peninsula (Kim et al., 2018)(*medium confidence*). Schofield et al. (2017) report a five-fold range of interannual variability in water column-integrated chlorophyll stocks, overlaid with a significant positive increase in the seasonal mixed-layer chlorophyll inventory over the twenty years of observations from the Palmer Long Term Ecological Research Station off the West Antarctic Peninsula (*low confidence*). The phenology of Southern Ocean phytoplankton blooms in this region may also be trending towards earlier in the growth season (Arrigo et al., 2017b)(*low confidence*). However, as highlighted in AR5 (Pörtner et al., 2014), the effect of climate change on Southern Ocean primary production is difficult to determine given that the length of time series data is insufficient (less than 30 years) to enable the climate change signature to be detected and attributed; and that, even when records are of sufficient length, data trends are often reported as being driven by climate change when they are due to a combination of climate change and variability (*very likely*).

Table 3.2: Model projections of trends due to climate-change driven alteration of phytoplankton properties under RCP8.5 from 2006–2100 across three zones of the Southern Ocean, modified from Leung et al. (2015). *Note that in many regions the zone between 40-50S represents the boundary between subtropical and subantarctic waters. [#]Temperature was not highlighted in the Abstract but reported elsewhere by Leung et al. (2015). Acidification was not reported as an important driver in this modelling experiment.

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

Model projections of trends in primary production in the Southern Ocean due to climate change are summarized in Table 3.2. Some of these model projections are supported by findings from manipulation studies conducted on Southern Ocean subpolar and polar phytoplankton species. Temperature is reported to play a central role in enhancing phytoplankton growth rates by 2100 (in tandem with increased iron) in polar waters (Xu et al., 2014; Hutchins and Boyd, 2016) (*medium confidence; low evidence*). In contrast to the model findings, Boyd et al. (2015) carried out experiments (light, temperature, CO_2, iron and nutrients) on a subantarctic pennate diatom and reported a doubling of growth rates was primarily driven by temperature, followed by iron supply, with little effect of ocean acidification (*low confidence*). The experiments conducted by Xu et al. (2014) on iron, light, CO_2, temperature in polar waters reveal a shift towards diatoms, and also decreases in cell size in both diatoms and *Phaeocystis* (fewer colonies and more uni-cells) (*low confidence, low evidence*). Recent studies on coastal phytoplankton indicate a detrimental effect of acidification (Hancock et al., 2017; Deppeler et al., 2018; Westwood et al., 2018) (*medium confidence, low evidence*). McMinn (2017) reviewed the effects of acidification on sea-ice algae, during laboratory manipulations lasting days to weeks, and reported that in general acidification caused no detrimental effects to the study organisms. In situ experiments also revealed a tolerance to acidification, and as for the laboratory studies provided evidence of either no change in metabolic rates or increased rates (*medium agreement, medium evidence*).

3.3.3.2.2 Antarctic krill and Southern Ocean microzooplankton

Previously reported declines in Antarctic krill abundance in the South Atlantic sector (Larsen et al., 2014a AR5, Chapter 28; Loeb and Santora, 2015) may reflect a step changes following an episodic period of anomalous peak abundance for this species (Fielding et al., 2014) rather than an ongoing decline (*medium confidence*). Recent analyses have not detected trends in long-term krill abundance in the South Atlantic sector (Larsen et al., 2014a; Kinzey et al., 2015; Steinberg et al., 2015). Nevertheless, as emphasized by Piñones and Fedorov (2016) in AR5 and given its dependence on sea ice habitats, the Antarctic krill population may already have changed and will be subject to further alterations (*high confidence*).

Predicted impacts of climate change on Antarctic krill relate primarily to changes in distribution as a consequence of changes in location of the optimum conditions for krill growth and recruitment (Melbourne-Thomas et al., 2016; Meyer et al., 2017; Suprenand and Ainsworth, 2017; Klein et al., 2018; Trebilco et al., In review). Based on empirical evidence for the relationship between temperature and krill growth and recruitment, the optimum conditions for krill are predicted to move polewards, with the decreases most apparent in the areas with the most rapid warming (*medium confidence*) (Section 3.3.1.2.1). The predicted impacts of temperature changes and ocean acidification on Antarctic krill are not homogeneously distributed; the greatest reductions in krill are predicted for the southwest Atlantic/Weddell Sea region (*low confidence*), which is the area of highest current krill concentrations, contains important foraging grounds for predators, and is also the area of operation of the krill fishery. 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 2010-2050 indicate a decline in krill biomass with contemporaneous increases in the biomass of gelatinous salps (Tarling et al., 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 1926–1938 and 1996–2013 showed no evidence of change despite a significant surface warming of 0.74°C (Steinberg et al., 2015)(*medium confidence*). These results suggest that predictions of distributional shifts based on temperature niches may not reflect the actual levels of thermal resilience of key taxa. Sub-decadal cycles of macrozooplankton community composition adjacent to the West Antarctic Peninsula are strongly linked to climate indices, with evidence of directional trends for some species over the period from 1993-2013, which may affect energy transfer to higher trophic levels and alter biogeochemical cycling (Manno et al., 2016)(*low confidence*). Pteropods are vulnerable to effects of acidification, and new evidence indicates that eggs released at high pCO_2 lack resilience to ocean acidification in the Scotia Sea region (Mintenbeck et al., 2012)(*medium confidence*).

3.3.3.2.3 Southern Ocean fish

Increasing water temperatures may displace notothenioid fishes of the family Channichthyidae in marginal habitats as they lack haemoglobin (an adaptation to cold temperatures) and are unable to adjust blood parameters to an increasing oxygen demand (Mintenbeck et al., 2012)(*low confidence*). Antarctic silverfish (*Pleurogramma antarctica*) are an important prey species in some regions of the Southern Ocean, and have an ice-dependent life cycle (Mintenbeck and Torres, 2017). Documented declines in the abundance of this species in some parts of the West Antarctic Peninsula may have consequences for associated food webs (Larsen et al., 2014a; Parker et al., 2015)(*low confidence*).

Myctophids and toothfish are important fish groups from both a food web (myctophids) and fishery (toothfish) perspective. A southward movement of isotherms in the Southern Ocean (Section 3.3.1.3.4) is expected to cause southward shifts in the distributions of myctophid fish species and could also result in isolated populations restricted to island shelves becoming locally extinct, if they are unable to adapt to warmer ocean temperatures (Larsen et al., 2014a)(*low confidence*). Postlarval toothfish are generalist predators that can migrate over large distances and occupy a very broad range of depths; hence, toothfish might be relatively resilient to environmental change by being able to descend or move to more favourable areas (Bost et al., 2009)(*low confidence*).

3.3.3.2.4 Southern Ocean seabirds and marine mammals

The distribution in time and space of marine mammals and seabirds has been associated with suitability of breeding habitats and with environmental features that facilitate the aggregation of prey (*medium confidence*)

(Bost et al., 2015; Kavanaugh et al., 2015; Hindell et al., 2016; Santora et al., 2017). The predictability of foraging grounds and ice-coverage are associated with variations in climate (Crocker et al., 2006; Baez et al., 2011; Dugger et al., 2014; Abrahms et al., 2017; Youngflesh et al., 2017) and are the main drivers of observed population changes of Southern Ocean (SO) higher predators (*high confidence*) (Ancona and Drummond, 2013; Ducklow et al., 2013; Chambers et al., 2014; Larsen et al., 2014a; Lyver et al., 2014; Bost et al., 2015; 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; Youngflesh et al., 2017). Biological parameters (reproductive success, mortality, fecundity, condition), life history traits, morphological, physiological and behavioural characteristics of species as well as processes/activities (migration, distribution, foraging, reproduction) are likely to change with changing climate as reported for marine birds, seals and whales (*high confidence*) (Whitehead et al., 2015; Braithwaite et al., 2015a; Seyboth et al., 2016; Hinke et al., 2017a). Thus, population trends for higher predators vary within and among Southern Ocean sectors (as defined by (Bost et al., 2009; Gutt et al., 2015; Hunt et al., 2016a; Murphy et al., 2016; Gutt et al., 2017; Trebilco et al., In review)) and reflect the different drivers affecting them, particularly sea-ice extent and food availability (*high confidence)* across regions (Section 3.3.1.3.4).

Gentoo penguin population estimates have increased (Lynch et al., 2013; Dunn et al., 2016; Hinke et al., 2017a), while some Adélie and Chinstrap penguin populations are reported to have declined (Trivelpiece et al., 2011; LaRue et al., 2013; Southwell et al., 2015; Cimino et al., 2016) (*high confidence*). Yet Youngflesh et al. (2017) suggest that populations shifts observed in Adelie penguin populations were more likely a result of strong phenological mismatch, i.e. interannual variability in good and bad years for prey and breeding habitat, than climate-change driven (Figure 3.11) (*medium evidence*). These authors found no evidence for directional change in phenological mismatch (*medium confidence*).

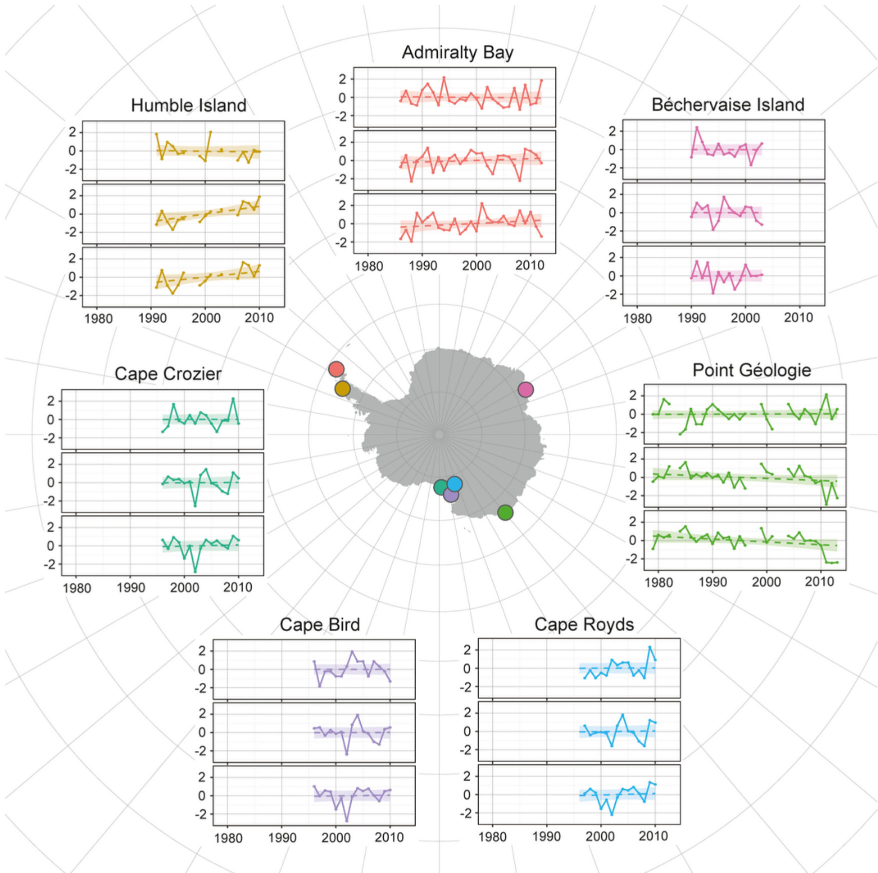


Figure 3.11: Time series for normalized Penguin breeding phenology (top panels), bloom mismatch index (middle panels), and sea-ice mismatch index (bottom panels) for each study site. Dashed lines represent model fit. Credible intervals (95%) are denoted by the shaded regions in each plot. Note that year t represents the austral summer spanning years t and t + 1. Site locations are represented on the map as coloured dots. From Youngflesh et al. (2017)

King penguin (Bost et al., 2015) and Emperor Penguin (Jenouvrier et al., 2014; Younger et al., 2015) populations are also declining throughout their range exhibiting lower foraging success and survival in relation to reductions in seasonal sea ice duration and longer/further foraging trips *(medium confidence*, *very likely*). Nevertheless, new evidence has suggested that present 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 emperor penguin colonies (Kooyman and Ponganis, 2017).

Evidence for climate change impacts on Antarctic flying birds remains limited (due to a low number of studies). Jenouvrier et al. (2015) report that contraction of sea ice near Terre de Adélie, East Antarctica has affected fledgling body condition and reduced the breeding success and population growth rates of Southern Fulmars. New evidence also indicates that increases in sea surface temperatures has led to reduced population growth for black-browed albatross (Pardo et al., 2017).

For SO marine mammals local and regional-scale oceanographic features and bathymetry that control prey aggregations both locally but also regionally will affect their ecological responses and biological traits (*high confidence*) (Lyver et al., 2014; Bost et al., 2015; Jenouvrier et al., 2015; Whitehead et al., 2015; Cimino et al., 2016; Seyboth et al., 2016; Hinke et al., 2017a; Pardo et al., 2017) and explain most of observed population shifts of marine mammals in the Southern Ocean (*likely*) (Kavanaugh et al., 2015; Hindell et al., 2016; Gurarie et al., 2017; Santora et al., 2017). Southern elephant seals (SES) in the Indian Sector of the SO have increased access to mesopelagic prey associated with decadal climate cycles but breeding SES females are excluded from highly productive continental shelf waters in years of increased sea ice extent and duration (*medium confidence)*(Hindell et al., 2016). Unlike SES, to date, there is no unified precise global estimate of the abundance of Antarctic pack ice seal species (Constable et al., 2017), even though this is essential to monitor changes in abundance of these ice-obligated predators in light of global warming and sea-ice change. Analysis of long-term data have suggested 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 SO as 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*). These changes reflect the interconnection of the effect of climate change on environmental conditions in foraging grounds (in the Southern Ocean) and in their breeding grounds (lower latitudes).

3.3.3.2.5 Southern Ocean pelagic and benthic ecosystem dynamics

Recent syntheses of Southern Ocean ecosystem structure and function recognise the importance of at least two alternative 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 (*medium confidence*) (Murphy et al., 2013; Murphy et al., 2016; Klein et al., 2018). Using an ecosystem model, Hauquier et al. (2016) found that the effects of warming on krill growth off the Antarctic Peninsula and in the Scotia Sea translated to increased risks of predator populations, particularly penguins, declining below a depletion threshold (75% of unimpacted levels) 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 in the Southern Ocean.

Ice-shelf retreat or collapse in Antarctica will lead to new marine habitats and to biological colonization (*high confidence*) (Figure 3.12). The loss of ice shelves and retreat of coastal glaciers around the Antarctic Peninsula 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 (Trathan et al., 2013), and have resulted in approximately 900×10^3 tonnes of new carbon uptake per year (Trathan et al., 2013)(*medium confidence*). New available habitat on coastlines may also afford breeding or haul-out sites for land based predators such as penguins and seals (Grange and Smith, 2013) (*low confidence*). Fjords that have been studied in the subpolar West Antarctic Peninsula are hotspots of benthic abundance and biodiversity (Grange and Smith, 2013) and there is evidence that glacier retreat in these environments can impact the structure and function of benthic communities (Moon et al., 2015) (*low confidence*). Such fjords also provide habitat and foraging areas for Antarctic krill and baleen whales, such that future change in these habitats may also impact pelagic ecosystems (Grange and Smith, 2013) (*low confidence*).

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

Polar zoobenthos blue carbon storage is predicted to increase with sea ice losses, because across-shelf growth gains from longer algal blooms outweigh ice scour mortality in the shallows (Clark et al., 2015)(*low confidence*). Communities in shallow water habitats mostly consist of dark-adapted invertebrates, and rely on sea ice to create low-light marine environments. Increases in the amount of light reaching shallow seabed under climate change may result in ecological regime shifts, in which invertebrate-dominated communities are replaced by macroalgal beds (Ingvaldsen and Gjøsæter, 2013; Clark et al., 2017)(*low confidence*).

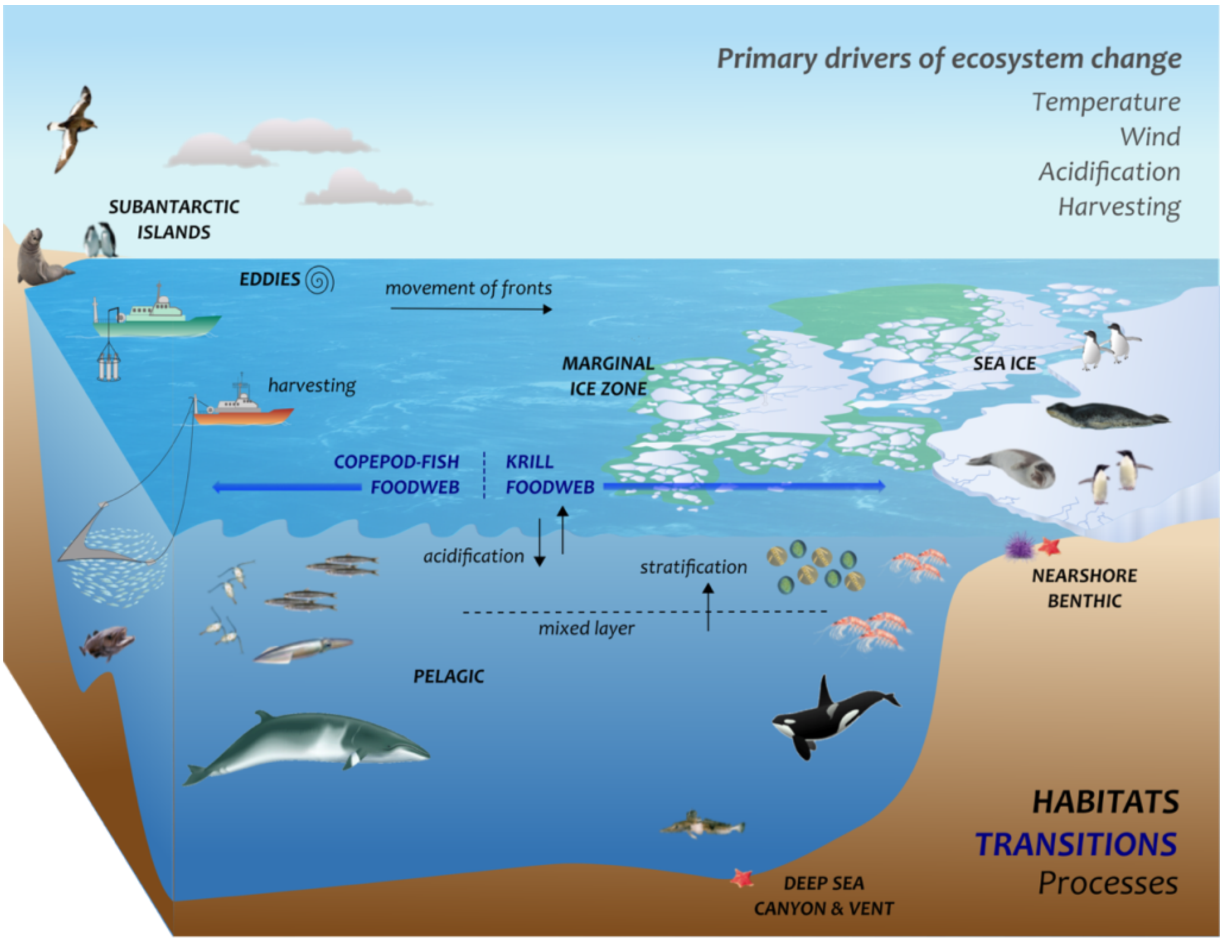


Figure 3.12: Schematic summary of for climate change effects on pelagic ecosystem structure and function across different habitat types in the Southern Ocean.

[START BOX 3.3 HERE]

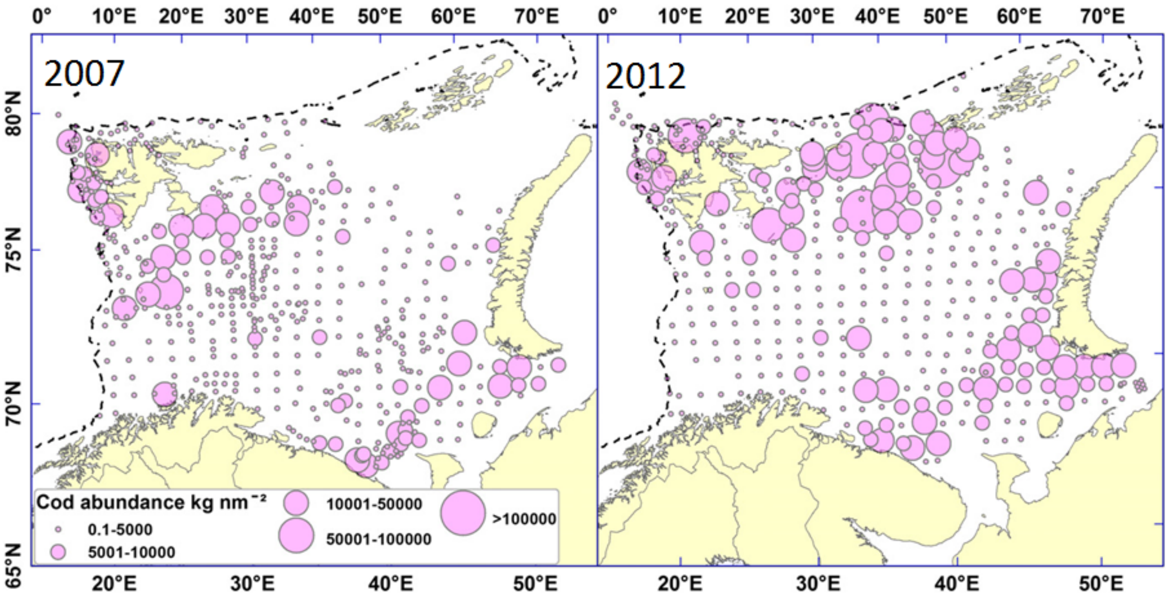
Box 3.3: The Implications of Climate-related Ocean and Cryosphere Change for Range Shifts and Invasive Species

Observations from recent surveys and laboratory studies confirm previous findings that a number of fish species in Arctic areas have changed their spatial distribution patterns substantially over the recent decades (*high confidence*). This may represent the establishment or extinction of populations in areas that are (now) environmentally suitable or unsuitable.

In the Barents and Bering Seas, recent changes in ocean conditions has impacted the summer feeding distributions of several pelagic and demersal fish species (*high confidence*). While some of the distributional responses are linked to climate change, in other cases the connection is more unclear because fish are responding to multiple stressors and in some case increasing population size. The summer feeding movements are impacted by a combination of factors including changes in suitable habitat availability (Kjesbu et al., 2014; Duffy-Anderson et al., 2017; Eriksen et al., 2017), prey availability, quality, and detection (Varpe et al., 2015; Hunt et al., 2016b), the presence of, and consumption by predators (Ingvaldsen and Gjøsæter, 2013), and population density (Kotwicki and Lauth, 2013) . The sensitivity of some fish species to these multiple stressors differs by size (Kjesbu et al., 2014; Barbeaux and Hollowed, 2018). Comparison of ocean conditions in the late 1970s and early 1980s with present conditions (2004), in the Barents Sea showed that suitable habitat for two abundant demersal fish stocks (Atlantic cod, *Gadus morhua*, and Northeast Arctic haddock, *Melanogrammus aeglefinus*) extended markedly to the north and east in response to increased sea temperature and retreating sea ice (Ingvaldsen and Gjøsæter, 2013; Landa et al., 2014; Eriksen et al., 2017) (Box 3.3, Figure 1). Similar shifts were also observed in pelagic species, with BS capelin (*Mallotus villosus*) shifting northwards in recent years (Ingvaldsen and Gjøsæter, 2013). The latter is probably an indirect climate effect, connected to the link between temperature and zooplankton production and distribution (Nøttestad et al., 2016). Comparison of the distribution of two dominant sub-arctic groundfish in the eastern Bering Sea (walleye pollock, *Gadus chalcogrammus*, and Pacific cod, *Gadus macrocephalus*) in cold (2011) and warm (2017) years revealed both species were distributed farther north in warm years.

The Northeast Atlantic stock of Atlantic mackerel (*Scomber scombrus*) has from 2007–2014 more than doubled its geographic range during the summer feeding season, mainly northwestwards into Icelandic and Greenland waters (Jansen et al., 2016; Nøttestad et al., 2016). In September 2013 mackerel were for the first time recorded in the Svalbard Archipelago. At above 78°N this is the northernmost known occurrence of mackerel. This expansion of distributional range is interpreted to be a result of the continued warming of the ocean in the region (Berge et al., 2015) (*high confidence*). Jansen et al. (2016), applying projections under RCP 4.5 and 8.5, suggest that the future warmer conditions in Greenland waters will allow the mackerel habitat to expand even further. Another species recently found in warming east Greenland waters, beyond its usual range, is Bluefin tuna (*Thunnus thynnus*), which has mackerel as a preferred prey species (MacKenzie et al., 2014). Evidence of climate driven spatial shifts in spawning locations is more limited, but has been described for Barents Sea cod (Sundby and Nakken, 2008). Spawn timing and locations have evolved over time in response to factors conducive to survival of young (Höffle et al., 2014; Kvile et al., 2017). This process is complicated with some species responding to local conditions and other species responding to regional ocean conditions (Stevenson and Lauth, 2012; Höffle et al., 2014).

Data on marine fish is severely limited in most other Arctic Ocean shelf regions. Several baseline studies were conducted in response to the International Polar Year (IPY) and several national expeditions. Results from these surveys reveal a latitudinal cline in the abundance of commercially harvestable fish species (Stevenson and Lauth, 2012). Observations of low level catches of sub-arctic species in the Chukchi and Beaufort Seas provides potential evidence of range expansions of some sub-arctic species (Logerwell et al., 2015). These limited surveys revealed that ocean conditions structure local distribution of fish (De Robertis et al., 2017). The spatial distribution of mid-water species also shows evidence of latitudinal partitioning between the four dominant species (Polar cod, *Boreogadus saida*, saffron cod, *Eleginus gracilis*, capelin, and Pacific herring, *Clupea pallasii*), with Polar cod being most abundant to the north (Bender et al., 2016).



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

Ongoing habitat degradation for ice-affiliated Arctic endemic marine mammals is an escalating threat, which is likely complicated by the northward expansion of the summer ranges of a variety of temperate species and increasing pressure from anthropogenic activities. Northward expansions of several whale species have been documented recently in both the Pacific and Atlantic sides of the Arctic (Brower et al., 2017) and most Arctic States are under increasing pressure to allow the expansion of Arctic tourism, northern fisheries, oil exploration and extraction, which creates increased ocean noise and other new risks to Arctic marine mammals (Zeller et al., 2011; Reeves et al., 2014; Thomas et al., 2016).

For the Antarctic, the increasing ice-free area on land linked to glacier retreat is expected to increase area available for new terrestrial ecosystems, and, along with growing tourist and science visitor numbers (Chown et al., 2017; Lee et al., 2017a), is expected to result in an increase in the establishment probability of terrestrial alien species (Hughes et al., 2015)(*medium confidence*). The number of such species along the Antarctic Peninsula is growing and forecast to continue to do so (Griffiths et al., 2013; Duffy et al., 2017)(*medium confidence*). For marine systems, introductions are expected to increase, especially of shell-crushing crabs, though evidence remains contested (Aronson et al., 2015; Tarling et al., 2017)(*very low confidence*). However a recent study of macrozooplankton assemblages in the Atlantic sector between 1926–1938 and 1996–2013 showed no evidence of change despite a significant surface warming of 0.74°C (CCAMLR, 1982)(*medium confidence*), suggesting that predictions of distributional shifts based on temperature niches may not reflect the actual levels of thermal resilience of key taxa.

Alien species (non-native species), in the context of Chown et al. (2012) are a major driver of local/global biodiversity change and ultimately loss. ‘Annex II of the Protocol on Environmental Protection to the Antarctic Treaty’ prohibits the introduction of non-native species to Antarctica as do the management authorities of sub-Antarctic islands (see De Villiers et al. (2006)). Despite this, alien species and their propagules continue to be introduced to the Antarctic and sub-Antarctic islands either via anthropogenic or natural means (Gutt et al., 2015; Houghton et al., 2016). The return to the wild of rehabilitated individuals has increased the potential for the propagation of non-native (invasive/alien) species into polar regions as defined by Chown et al. (2017). The recent sightings of California gray whales in the Atlantic and Mediterranean show the magnitude of movement by migratory individuals and potential propagation of zoonoses given their traditional poleward specific migrations (Pacifici et al., 2015; Pauchard et al., 2016).

[END BOX 3.3 HERE]

3.3.4 Sectoral Consequences of Changing Polar Oceans and Sea Ice

3.3.4.1 Fisheries

3.3.4.1.1 Arctic fisheries

Descriptions of Arctic fisheries and their responses is discussed on Section 3.5.5.2, here we discuss the potential impacts of climate change on marine fisheries. Seasonal and interannual variability in ocean conditions is expected to influence product quality, quantity and catchability (Haynie and Pfeiffer, 2012). As documented in Section 3.3.3, under most RCPs climate change will affect the spatial distribution and productivity of some marine fish and shellfish. The magnitude of these changes differ across species or stocks depending on the vulnerability of the species to changing environmental conditions. These spatial shifts will impact community access to fish, the costs of fishing and transboundary fishing agreements (Table 3.5). As also documented in Section 3.3.3 climate change influences the boundary between subarctic and high Arctic fish communities, with unclear effects on future fisheries.

The existence of science based holistic management strategies in the southeastern Bering Sea portends that the management of marine resources in the Arctic may be founded in precautionary approaches to sustaining marine resources and ecosystem structure to the extent possible.

If managed sustainably, Arctic fisheries may be able to adapt to moderate future warming (European Parliament's Committee on Fisheries, 2015). For example, the Norwegian cod fishery has exported dried cod over an unbroken period of more than thousand years (Barrett et al., 2011), reflecting the resilience of the northern Norwegian cod fisheries to historic climate variability (Eide, 2017). The high present yield of the Barents Sea (Section 3.3.3.1) and model projections indicate that enlarged habitat and increased production of plankton and prey due to increasing temperatures and ice retreat, may ensure that the migratory fish stocks remain large and the economic benefits from fisheries continues (Lam et al., 2016; Eide, 2017).

Five nations have existing EEZs in the high Arctic and each nation manages their resources within the regulatory measures of their nation. Commercial fishing is currently prohibited in the US portions of the Chukchi and Beaufort Seas (Wilson and Ormseth, 2009). In the Canadian sector of the Beaufort Sea commercial fisheries is until now only small scale and locally operated, but climate change with decreasing ice cover together with over-harvesting of fish stocks other places may increase the incitement. This has caused concern among local Inuvialuit subsistence fishers and a new proactive ecosystem-based Fisheries Management Framework was developed (Ayles et al., 2016). In 2015, the Oslo declaration on high seas fishing in the central Arctic Ocean was signed which established a moratorium on commercial fishing in the central Arctic Ocean and encouraged research cooperation amongst the bordering nations. These constraints will limit the expansion of commercial fishing until sufficient information if available to sustainably manage fisheries under the influence of climate change.

A review of future harvest of living resources in the European Arctic by Haug et al. (2017) points towards high probability of increased northern movement of several commercial fish species (see Section 3.3.3.1 and Box 3.3), but only to the shelf slope for the demersal species. This suggests increased northern fishing activity, but within the 200 nm zones and the present management system (Haug et al., 2017).

3.3.4.1.2 Antarctic fisheries

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 tootfish combined (CCAMLR, 2017a). The fishery for Antarctic krill in the southern Atlantic sector and the northern West Antarctic Peninsula (together the current area of focus for the fishery) has become increasingly concentrated in space over recent decades, which has raised concern regarding localised impacts on krill predator (Hinke et al., 2017a). The krill fishery has also changed its peak season of operation. In the early years of the fishery, most krill were taken in summer and autumn, with lowest catches being taken in spring. In recent years krill catches have shown a reversal in this historic trend with lowest catches occurring over summer, peaking in late autumn, with very little fishing activity in spring (Nicol and Foster, 2016). Some of these temporal and spatial shifts in the fishery over time have been attributed to reductions in winter sea-ice extent in the region (Kawaguchi et al., 2009) (*medium confidence*). Recent increases in the use of krill catch to produce krill oil (as a human health supplement) has also led to vessels concentrating on fishing in autumn and winter when krill are richest in lipids (Nicol and Foster, 2016)(*medium confidence*).

3.3.4.2 Tourism

Reductions in sea ice related to climate change has facilitated an increase in marine and cruise tourism opportunities across the Polar Regions related directly to an increase in open water areas and season lengths (*very high confidence*) (Dawson et al., 2014; Dawson et al., 2017). Alaska attracts the highest number of passengers annually at just over one million; Svalbard attracts 40,000 to 50,000; Greenland 20,000-30,000; and Canada 5,000-8,000 (Dawson et al., 2017). Over the past decade, Arctic cruise traffic has increased, ships are traveling further in a single season, more itineraries are being sold, larger vessels with more passenger berths are in operation, purpose-built polar cruise vessels are being constructed, and smaller private yachts are appearing in greater frequency (Lasserre and Têtu, 2015; Johnston et al., 2017; Dawson et al., 2018). Compared to other forms of maritime traffic (i.e. re-supply or trade), the cruise tourism industry is highly adaptable and can respond quickly to take advantage of opportunities emerging because of climate change.

Observed increases in cruise and yacht traffic in the Polar Regions is linked to both 'current' and 'anticipated' climate change-related opportunities and perceptions. For example, Canada's Northwest Passage (southern route) is now reliably accessible during the summer cruising season and as a direct result of observed reductions in sea ice, tourism operators now regularly offer Northwest Passage itineraries, which has resulted in a 70% increase in passenger vessel traffic and a 400% increase in pleasure craft in that area since 2012 (Johnston et al., 2017; Dawson et al., 2018). The anticipated implications of future climate change have led to the emergence of a niche tourism market being called 'last chance tourism' – whereby tourists explicitly seek to experience vanishing landscapes or seascapes, and disappearing natural and social heritage before they disappear (Lamers et al., 2013). There is *high confidence* that polar cruise tourism will continue to grow over the coming decade (Johnston et al., 2017).

Increases in Polar cruise tourism have important risks and opportunities related to employment, development, education, safety, security, and environmental and cultural 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). There are also important international scientific opportunities related to the use of polar cruise vessels as 'ships of opportunity' where vessels can be equipped with instrumentation. Because the sector relies on a set of regulations that apply to all types of maritime shipping, yet cruise ships purposefully travel off regular shipping corridors, a need for appropriate governance regimes, specialized infrastructure, and focused policy attention has been identified (Dawson et al., 2014; Pashkevich et al., 2015; Dawson et al., 2016; Dawson et al., 2017). Private pleasure craft (yachts) remain almost completely unregulated and pose unique risks in the future (Johnston et al., 2017).

Almost 37,000 predominantly ship-borne tourists visited Antarctica in 2016-17 (IAATO, 2017). It is expected these numbers will rise (IAATO, 2017). Due to accessibility and convenience, these tourism operations are mostly based around the few ice-free areas of Antarctica, and are concentrated on the Antarctic Peninsula (Pertierra et al., 2017). The biodiversity supported by ice-free areas, particularly those on the Antarctic Peninsula, has been identified as being particularly vulnerable to the changing climate (Tin et al., 2013) and to the introduction of terrestrial alien species (Hughes et al., 2015; Duffy et al., 2017; Lee et al., 2017a)(*medium confidence*).

3.3.4.3 Transportation

The Arctic is reliant on marine transportation for the import of food, fuel, and other goods. 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 also increasing as the region becomes more accessible because of reduced sea ice cover. There are four major Arctic international trade routes: the Northwest Passage (NWP), the Northern Sea Route (NSR), Arctic Bridge (AB), and Transpolar Sea Route (TSR). All of these routes offer significant trade benefits as they provide substantial distance savings compared to traditional international trade routes via the Suez or Panama Canal. However, variable and dynamic ice conditions currently limit trade through many of these routes (in particular the NWP) where distance savings are not translated into time savings because of the challenges involved in navigating ice-infested waters (Smith and Stephenson, 2013).

There is *very high confidence* in an observed increase in Arctic shipping activity over the past decade (Pizzolato et al., 2014; Eguíluz et al., 2016; Pizzolato et al., 2016; Dawson et al., 2018). While there are many relevant factors such as natural resource development, regional trade, geopolitics, commodity prices, global economic and social trends, national priorities, tourism demand, ship building technologies, and insurance costs (Lasserre and Pelletier, 2011; Têtu et al., 2015; Dawson et al., 2017), there is *high confidence* that the relative strength of climate change driven reductions in sea ice as a driver of shipping change has increased in recent years (Dawson et al., 2017).

Increased shipping activity across the Arctic is projected for the future as northern routes become increasing accessible due to climate change and as insurance companies respond to decreased risks related to Polar transport. The NSR is expected to be more viable than others, considering investments in infrastructure and favourable sea ice dynamics. In comparison, the NWP and AB have limited port and marine transportation infrastructure, limited soundings and incomplete hydrographic charting, and challenging sea ice conditions (Stephenson et al., 2013; Andrews et al., 2018). These conditions, together with limited search and rescue capacity and remote and harsh geography compound risks from shipping activity across the entire region (Dawson et al., 2017). Recent studies suggest navigation will likely become easier and faster in the Arctic by the mid-21st century (Stephenson et al., 2011; Smith and Stephenson, 2013; Barnhart et al., 2015; Melia et al., 2016). These projected changes to Arctic shipping activities will have significant socio-economic and political implications related to safety (i.e. marine accidents, local accidents, ice as a hazard), security (i.e. trafficking, terrorism, etc.), and environmental and cultural sustainability (i.e. invasive species, marine mammals, Arctic shore birds, impacts to subsistence hunting, etc.) (Arctic Council, 2015).

The predominant shipping activity in Antarctica is logistic in nature, supporting land-based stations; or marine research vessels. Changing sea ice conditions and duration have presented significant challenges to this shipping activity (Lieser et al., 2013; Chown, 2017) (*medium confidence*). Navigational hazards from less predictable sea ice conditions also increases the chances of vessels becoming trapped in the ice, leading to safety and environmental risks.

3.4 Changing Polar Seasonal Snow Cover, Permafrost and Freshwater Ice: Global and Local Impacts

3.4.1 Observations of the Terrestrial Cryosphere

3.4.1.1 Seasonal Snow Cover

Snow cover is a defining characteristic of the Arctic land surface for up to 9 months each year. Changes in seasonal terrestrial snow cover exert important influences on the surface energy budget, ground thermal regime, and freshwater budget of the Arctic. Snow cover also interacts with vegetation, influences biogeochemical activity, with consequences for ecosystem services.

3.4.1.1.1 Snow cover extent/snow cover duration

The Arctic land areas north of 60°N are always completely snow covered in winter, so the transition seasons of fall and spring are key when characterizing variability and change. Dramatic reductions in Arctic spring snow cover extent (SCE) have occurred since satellite charting began in 1967 (Estilow et al., 2015), particularly since 2005, with the rate of change (1967-2017) in May and June (the months of maximum snow retreat) of -5.0% and -17.8% per decade respectively (Derksen et al., 2017). An evaluation of SCE trends using multiple data sources (surface observations, passive microwave satellite data, land surface models driven by reanalysis) covering different time periods shows there is *high confidence* in the magnitude of May SCE reductions, but only *medium confidence* in June trends because of a large inter-dataset spread (-5% to -16% per decade) (Hori et al., 2017; Mudryk et al., 2017a).

There is *very high confidence* that the loss of spring SCE is also reflected in shorter snow cover duration (SCD; reduced SCE = earlier loss of snow and hence shorter SCD) trends during the spring derived from surface observations (Bulygina et al., 2011; Brown et al., 2017a), satellite data (Wang et al., 2013; Estilow et al., 2015), and model-based analyses (Liston and Hiemstra, 2011). These SCD trends range between –0.7 and –3.9 days per decade depending on region and time period, but all spring SCD trends from all datasets are negative (Brown et al., 2017a). These same multi-source datasets also identify reductions in fall SCE and SCD (–0.6 to –1.4 days per decade; summarized in Brown et al. (2017a)) (*very high confidence*). Notable outliers are positive trends in October and November SCE in the NOAA-CDR (Hernández-Henríquez et al., 2015) which are not replicated in other surface, satellite, and model datasets (Brown and Derksen, 2013; Mudryk et al., 2014). SCE trend sensitivity to surface temperature forcing in the NOAA-CDR is anomalous compared to other datasets for these months (Mudryk et al., 2017a); there is *low confidence* in the NOAA-CDR trends for these two months.

3.4.1.1.2 Snow depth/snow water equivalent

There is emerging evidence of negative trends in pre-melt maximum snow depth (SDmax) from weather station observations across the Russian Arctic over the 1966-2014 time period (Bulygina et al., 2011; Osokin and Sosnovsky, 2014). There is only *medium confidence* in these observations because trend differences between open and forested sites (Maksyutova et al., 2012; Brown et al., 2017a) illustrate that the pointwise nature of station snow depth measurements do not capture prevailing conditions across the landscape. SDmax trends over the North American Arctic are mixed and largely insignificant statistically (Vincent et al., 2015; Brown et al., 2017a). The timing of SDmax has shifted earlier by –2.7 days per decade for the North American Arctic (Brown et al., 2017a); comparable analysis is not available for Eurasia.

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

3.4.1.1.3 Drivers

Changes in seasonal terrestrial snow cover across Arctic land areas are driven by processes that influence the initial deposition of snow in the autumn, the evolving properties of the snowpack during the snow accumulation season, and the timing and intensity of the spring melt period. It is *virtually certain* that observed changes in Arctic SCE and SCD (Section 3.4.1.1.1) are linked to the observed surface temperature increases across the Arctic over recent decades (Hawkins and Sutton, 2012; Fyfe et al., 2013), which are amplified compared to global trends (Serreze et al., 2009; Fyfe et al., 2016; Overland et al., 2017a) because of a number of feedbacks which operate in polar regions (Pithan and Mauritsen, 2014) (see Section 3.A.1.2). Despite uncertainties due to sparse measurements (Cowtan and Way, 2014), observed warming across Arctic land areas has seasonal peaks with maxima in the fall and spring periods (Brown et al., 2017a) - this seasonal pattern shortens the length of the snow accumulation season (delayed snow onset in fall; summarized in Brown et al. (2017a), reduces the fraction of precipitation falling as snow (Screen and Simmonds, 2011), and initiates earlier melt in the spring (Brown and Derksen, 2013).

Based on multiple datasets/periods, there is *high confidence* in a consistent sensitivity for Arctic SCE of –0.8 to -1.0×10^6 km^2 K^{-1} for spring(Brown et al., 2010; Brown and Derksen, 2013) and –0.7 to -0.8×10^6 km^2 K^{-1} for fall(Derksen and Brown, 2012; Brown and Derksen, 2013). There is *very high confidence* that changes in Arctic SCE can be directly related to extratropical temperature increases, with sensitivities ranging between −2.5% and –3% K^{-1} (Brutel-Vuilmet et al., 2013; Thackeray et al., 2016; Mudryk et al., 2017a).

Snowfall-based drivers of Arctic snow cover changes are much more uncertain because precipitation remains a sparse and highly uncertain measurement over Arctic land areas. While efforts are underway to improve the coordinated correction of systematic precipitation measurement errors (Kochendorfer et al., 2017), in situ datasets remain uncertain (Yang, 2014) and largely regional (for example, Kononova (2012); (Vincent et al., 2015). Atmospheric reanalyses provide another perspective on Arctic precipitation (Vihma et al., 2016) but these products are inconsistent and remain poorly validated (Serreze et al., 2012). Previous assessments have identified positive trends in Arctic precipitation (Min et al., 2008; Callaghan et al., 2011; Hartmann et al., 2013) but there is little recent evidence of coherent snowfall trends across the Arctic.

While the initial snowfall event in autumn definitively sets the timing of snow onset on the land surface, winter season changes in snowfall have a more uncertain impact on Arctic snow depth and SWE because of the influence of wind redistribution. Arctic snow accumulates to the height of the prevailing ground vegetation, after which it is redistributed to spatially constrained topographic depressions and drifts with potentially high sublimation loss (Sturm and Stuefer, 2013). This loss has been identified as an important factor in alpine areas and on ice sheets (MacDonald et al., 2010; Lenaerts and van den Broeke, 2012) but remains a key uncertainty in the mass budget of the Arctic snowpack.

Regional scale changes in Arctic ground vegetation could be an important driver of changes in snow catchment across tundra regions: more expansive/taller shrubs result in greater snow catchment and retention with consequent snow/shrub feedbacks (as postulated in Sturm et al. (2001)). While the impacts of increased shrub cover on snow insulating properties and ground temperatures are well understood (Gouttevin et al., 2012) changes in vegetation cover across the Arctic (at the coherent regional scales needed to impose an impact on the hydro-climatic system) are not uniform and the drivers are poorly understood (Myers-Smith et al., 2015). Vegetation changes can also influence spring snow melt rates via changes to albedo (Marsh et al., 2010; Loranty et al., 2014).

The potential influence of changing Arctic sea ice conditions on seasonal terrestrial snow is an emerging area of research. Reanalyses and model simulations suggest a moistening of the Arctic atmosphere in response to reduced sea ice extent (Liu et al., 2012; Screen et al., 2013; Petrie et al., 2015). Temperature and snowfall departures over Eurasia have been statistically associated with regions of sea ice loss (Mori et al., 2014; Wegmann et al., 2015) but the circulation impacts and driving mechanisms remain uncertain (Li and Wu, 2012; Barnes and Screen, 2015) The observed increase in atmospheric moisture also results in enhanced longwave radiation, with feedbacks to snow melt (Wang et al., 2015) and land surface hydrological terms (Porter et al., 2012).

3.4.1.2 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), and northern populations (Takakura (2018); Section 3.4.3.3.1). Assessing these impacts requires consideration of complex inter-connected processes, some of which are only partly observed by surface networks. The increasing imprint of human development, such as flow regulation on major northerly flowing rivers adds complexity to the determination of climate-driven changes.

3.4.1.2.1 Freshwater ice

Confidence in lake ice cover trends from surface measurements across the Arctic are limited by exceptionally sparse monitoring networks, and short satellite time series. Between 1855 to 2005, reductions in total ice cover duration were observed in the northern European Arctic with earlier spring melt accounting for the majority of this change (Benson et al., 2012). This is similar to lakes in the Canadian Arctic (Prowse et al., 2011b). Long-term in situ river ice records indicate that the duration of ice cover in Russian Arctic rivers decreased by 7 to 20 days over 1955–2012 (Lammers and Shiklomanov, 2014); consistent a separate analysis covering 1979–2009 (Park et al., 2015) (*high confidence*).

Analysis of optical satellite imagery for 13,300 lakes larger than 1 km^2 between 2000-2013 identified a significant trend of earlier spring ice break-up across all regions of the Arctic, driven primarily by surface temperature (Šmejkalová et al., 2016). A recent circumpolar assessment based on the analysis of satellite passive microwave observations showed approximately 80% of Arctic lakes have experienced declines in ice cover duration during 2002–2015, due to both a later freeze-up and earlier break-up (Du et al., 2017), consistent with regional analyses of river ice (Cooley and Pavelsky, 2016; Pavelsky and Zarnetske, 2017) (*high confidence*). There are indications that lake ice across Alaska has begun thinning in recent decades (Alexeev et al., 2016), but ice thickness trends are not available at the pan-Arctic scale.

Analysis of satellite radar 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) (*high confidence*).

3.4.1.2.2 Surface water and runoff

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

Thermokarst lake expansion has been observed in the continuous permafrost of northern Siberia in both regional (Smith et al., 2005) and more localized studies (e.g., on the Yamal Peninsula, Sannikov (2012)) (*high confidence*). Net surface water area reduction has been observed in discontinuous permafrost of central and southern Siberia (Smith et al., 2005; Kirpotin et al., 2008; Sharonov, 2012), Canada (Labrecque et al., 2009; Carroll et al., 2011) and interior Alaska (Chen et al., 2012; Rover et al., 2012) (*high confidence*). One notable exception to observed surface water area trends is the multi-decadal loss of abundance and coverage of ponds in the Arctic coastal plain of Alaska with continuous permafrost coverage (Andresen and Lougheed, 2015). Increased evaporation from warmer, longer summers and increased transpiration from encroaching algal mats are potential factors contributing to pond reduction. In addition to permafrost thaw, hydroclimatic factors and other processes such ice-jam flooding (Chen et al., 2012; Jepsen et al., 2015) are important considerations for understanding the underlying mechanisms of observed surface water change across the Arctic.

A general trend of increasing discharge has been observed for large Siberian (Peterson et al., 2002; Troy et al., 2012) and Canadian (Ge et al., 2013; Déry et al., 2016) rivers that drain to the Arctic Ocean; extreme regional runoff events have also been identified (Stuefer et al., 2017). The magnitude of these trends (1976–2015) are 3.1 ± 2.0% for the Eurasian rivers and 2.6 ± 1.7% for the North American rivers (Holmes et al., 2015) (Figure 3.5). There is *high confidence* that the detected increase in baseflow in the North American Arctic (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009) and northern Eurasia (Smith et al., 2007a; Duan et al., 2017) over the last several decades is attributable to permafrost thaw and the concomitant enhancement in groundwater discharge. This discharge represents a notable heat flux to the Arctic Ocean (Yang et al., 2014), with impacts on coastal sea ice processes (i.e., Kuzyk et al., 2008). The timing of spring season peak flow is generally earlier (Ge et al., 2013; Holmes et al., 2015), largely attributed to earlier snow melt onset (Yang et al., 2007) (*very high confidence*). There is 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).

Analyses of long-term water temperature records shows significant increases attributed to climate warming (Webb et al., 2008; Yang and Peterson, 2017). There is uncertainty, however, in the extent to which rising air temperatures can be directly linked to water temperature in rivers because of the influence of reservoir regulation (Liu et al., 2005; Lammers et al., 2007). There is also high uncertainty in net organic carbon export to the Arctic Ocean due to limited historical data (Holmes et al., 2002) and non-uniform responses in carbon export to permafrost thaw observed in different watersheds (Tank et al., 2012; Vonk et al., 2015).

3.4.1.2.3 Drivers

Anthropogenically driven temperature increases across the Arctic are well documented (Fyfe et al., 2013; Overland et al., 2017a), and affect freshwater systems by driving variability and change in snow cover extent and duration (Section 3.4.1.1.1), freshwater ice thickness and duration (Section 3.4.1.2.1), river discharge (Section 3.4.1.2.2) and the ground thermal regime (Section 3.4.1.3). Additional drivers to changes in freshwater systems are precipitation (including moisture availability and phase), evapotranspiration, and land cover change.

Zhang et al. (2013) argue that increases in poleward atmospheric moisture transport (and hence precipitation) are responsible for observed increases in discharge from northern rivers into the Arctic Ocean (see Section 3.3.1.2.2). This is consistent with increased moisture storage in a warmer Arctic lower troposphere (Held and Soden, 2006). While a number of products suggest increases in Arctic precipitation in recent decades (Lique et al., 2016; Vihma et al., 2016), there is *low confidence* in reanalysis-based closure of the Arctic freshwater budget (i.e., Serreze et al., 2006) due to a wide spread between available reanalysis derived precipitation estimates (Lindsay et al., 2014). While temperature driven reductions in summer Arctic snowfall have been identified (Screen and Simmonds, 2011), there is no evidence (*high confidence*) of trends in rain-on-snow events, which can have important ecological implications (Cohen et al., 2015; Dolant et al., 2017).

Evapotranspiration (ET) is a poorly constrained observation across Arctic land areas. Studies suggest increases in ET of +0.13 to +0.38 mm/y/y using satellite (Zhang et al., 2009), and model-derived datasets (Rawlins et al., 2010; Liu et al., 2014; Liu et al., 2015a), but there is only *medium confidence* in this trend because time series are not current (typically ending before 2010) and updated Arctic ET assessments are sparse. Observed increases in temperature, enhanced precipitation, a shorter snow cover season coupled with a longer growing season (which are themselves linked; Yi et al. (2015); Pulliainen et al. (2017)) are all consistent with estimates of increased ET, but there is considerable uncertainty in each of these terms.

Landscape alterations, including disturbance and shifting vegetation patterns also play a key role in driving changes to freshwater systems (Wrona et al., 2016). In permafrost regions, surface elevation changes due to thaw subsidence in thermokarst-affected landscapes substantially drive hydrologic change by forming depressions for lake formation or generating lake drainage pathways (Jones et al., 2011; Grosse et al., 2013). In addition, the gradual increase of the seasonal active layer thickness in a warmer climate impacts temporary water storage and thus runoff regimes in drainage basins. Eventually, thermokarst-driven 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 (Rowland et al., 2011b). The extent to which vegetation changes have impacted pan-Arctic ET (in the context of greater water availability versus a longer growing season) is not well established (Bring et al., 2016), but vegetation changes influence other hydrological processes by impacting ground temperatures and permafrost (Nauta et al., 2014).

3.4.1.3 Permafrost ground

3.4.1.3.1 Temperature

Permafrost is perennially frozen ground (rock, soil, ice) that underlies natural ecosystems and human communities in high latitude and some high altitude areas of Earth. Permafrost is narrowly defined by temperature (< 0°C for at least 2 consecutive years), essentially marking the long-term phase change from liquid water to ice. Permafrost temperature, at a depth where seasonal temperature variation is negligible, provides an indication of long-term change in climate. In contrast, the thickness of the surface active layer, which thaws and re-freezes seasonally, records effects of shorter term fluctuations in climate on permafrost, in particular summer air temperature and precipitation. Continuing the trend from AR5, record high temperatures in the upper permafrost (~10-20 m depth) have been documented at all permafrost monitoring sites on the North Slope of Alaska and at most long-term monitoring sites in the Northern Hemisphere permafrost domain (*high confidence*)(AMAP, 2017b) (Figure 3.13). At some locations, the temperature is 2-3°C higher than 30 years ago. Since 2000, the typical rate of increase in permafrost temperatures was between 0.4°C and 0.7°C per decade for colder continuous permafrost monitoring sites and between 0.1°C and 0.2°C for warmer discontinuous permafrost. Relatively smaller increases in permafrost temperature in warmer sites indicate that permafrost is degrading as a consequence of the surface active layer increasing in thickness. In contrast to temperature, there is only *medium confidence* that active layer thickness is increasing, because decadal trends vary across regions and sites (Shiklomanov et al., 2012) and because mechanical probing of the thawed surface layer can underestimate the degradation of surface permafrost in some cases (Streletskiy et al., 2017). Site averages in three of six Arctic study regions (Russian Far East, Russian European North, East Siberia) show a decadal trend of increasing active layer thickness, whereas three other regions (West Siberia, North Slope Alaska, Northwest Canada) do not show this trend (Romanovsky et al., 2016a; AMAP, 2017b).

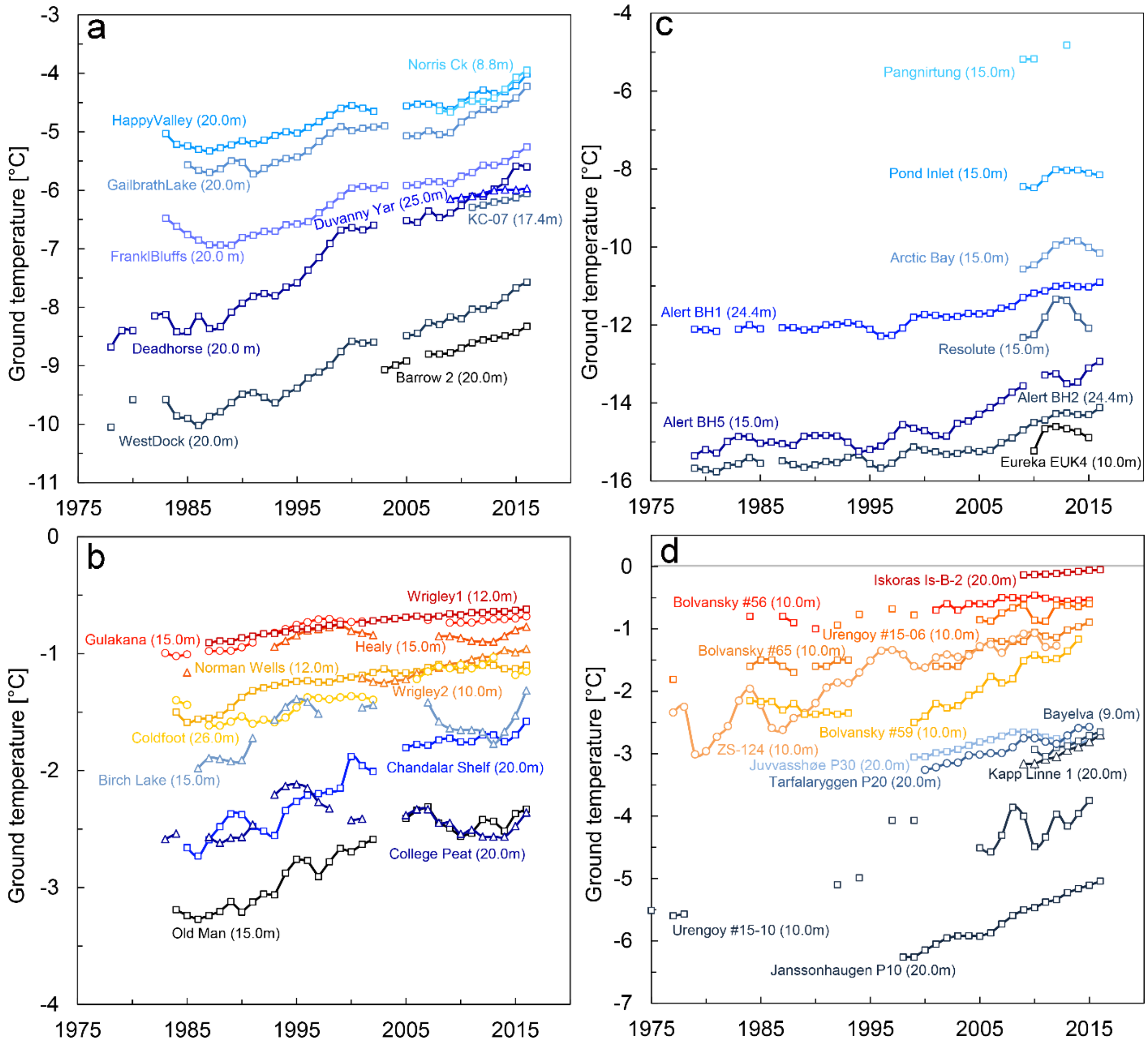


Figure 3.13: Time series of mean annual ground temperature at depths of 9 to 26 m below the surface at selected measurement sites across: a) cold continuous permafrost of NW North America and NE of East Siberia; b) discontinuous permafrost in Alaska and northwestern Canada; c) cold continuous permafrost of Eastern and High Arctic Canada; d) Continuous to discontinuous permafrost in Scandinavia, Svalbard, Russia/Siberia. Temperatures are measured at or near the depth of zero annual amplitude. (Data series updated from: Christiansen et al., 2010; Ednie and Smith, 2015; Smith et al., 2015b; Romanovsky et al., 2016b; Smith et al., 2016b; Boike et al., 2018).

3.4.1.3.2 Ground ice

The amount and distribution of ice within permafrost ground is controlled by the local and regional climate as well as by geomorphological processes over recent geologic time (French and Shur, 2010; Gilbert et al., 2016). On a very local scale (<m^2 scale), ground ice content can range from <10% in ice-poor permafrost up to 100% in the form of massive ice within the soil, in particular in fine-grained sediments and soils that restrict water drainage. Higher ice content slows the advance of permafrost thaw because of the substantial heat required to turn ice to liquid (Burn and Nelson, 2006). At the same time, thawing of high-ice permafrost has more serious consequences relative to low-ice permafrost. When conditions change and ground ice melts, the ground surface subsides and collapses into the volume previously occupied by ice, causing disturbance to overlying ecosystems and human infrastructure (Jorgenson et al., 2013). On a regional scale, ice content of frozen ground has been categorized into high (>20%), medium (10-20%), and low (<10%) ice content by volume (Zhang et al., 2000). Roughly two-thirds of the total permafrost zone is classified as having low ice content, with the other one-third having medium or high ice content (*low-medium* *confidence*). The high ice category has a large range of ice content, including the yedoma (Ice Complex) deposits in Siberia, Alaska, and the Yukon in Canada containing up to 50-80% ground ice by volume (Zimov et al., 2006; Schirrmeister et al., 2011; Strauss et al., 2017). Other regions including Northwestern Canada, the Yamal and Gydan peninsulas of West Siberia, and smaller portions in Eastern Siberia and Alaska regions contain buried glacial ice bodies of significant thickness and extent (Lantuit and Pollard, 2008; Leibman et al., 2011; Kokelj et al., 2017). Higher resolution ground ice maps exist in some regions where economic development resulted in engineering geological assessments of permafrost for planning purposes (Trochim et al., 2016; Vincent et al., 2017), but this resolution is still lacking at the pan-Arctic scale (Jorgenson and Grosse, 2016).

3.4.1.3.3 Carbon

Northern soils have long been known to contain large amounts of organic carbon, accumulating in frozen and waterlogged soils (Gorham, 1991). But only recently has there been focus on carbon stored deeper in permafrost soils (Zimov et al., 2006), below the traditional 1-meter accounting depth (Ping et al., 2008). Soils in the northern permafrost zone have unique characteristics that can cause an accumulation of deep carbon. These characteristics include: vertical mixing due to the freeze-thaw cycle, peat accumulation as a result of waterlogged conditions, and deposits of wind and water-moved sediment (yedoma / loess) tens of meters thick (Gorham, 1991; Schirrmeister et al., 2002; Bockheim and Hinkel, 2007; Schuur et al., 2008). Furthermore, there are marine permafrost carbon deposits today that formed via the same processes when sea level was lower during last glacial period. The permafrost zone represents a large, climate-sensitive reservoir of carbon with the potential to be rapidly decayed and transferred to the atmosphere as carbon dioxide or methane as permafrost thaws in a warming climate, thus accelerating the pace of climate change (Schuur et al., 2008; Schuur et al., 2015).

The current best estimate of total organic soil carbon (terrestrial) in the northern circumpolar permafrost zone ($17.8 x 10^6$ km^2 area) is 1460 to 1600 petagrams (*medium confidence*) (Pg; 1 Pg = 1 billion metric tons) (Hugelius et al., 2014; Schuur et al., 2015; Strauss et al., 2017). This inventory includes all soil orders within the permafrost zone and thus also counts carbon in nonpermafrost soil orders, active layer (surface) carbon that thaws seasonally, and peatlands. All permafrost zone soils estimated to 3 m in depth contain 1035 ± 150 PgC (Hugelius et al., 2014) (*high confidence*), with two-thirds of the soil carbon pool in Eurasia, and the remaining one-third in North America (including Greenland) (Tarnocai et al., 2009). Of this amount, 800–1000 billion tons is perennially frozen, with the remainder contained in seasonally thawed soils. The 1035 Pg of soil carbon quantified from the northern circumpolar permafrost zone adds another 50% to the global 3-m inventory (2050 Pg C, excluding tundra and boreal biomes (Jobbágy and Jackson, 2000)), even though it occupies only 15% of the total global soil area (Schuur et al., 2015).

Substantial permafrost carbon exists below 3 m depth (*medium confidence*). Deep carbon has been best quantified for the yedoma region of Siberia and Alaska, characterized by permafrost sediments tens of meters thick. The yedoma region covers a $1.4 x 10^6$ km^2 area that remained ice-free during the last Ice Age (Strauss et al., 2013). The carbon inventory of this region comprises yedoma soils that were previously thawed as lakes formed and then refrozen into permafrost when lakes drained, interspersed by intact permafrost yedoma deposits that were unaffected by thaw-lake cycles (Walter Anthony et al., 2014). Together, this region accounts for 327 to 466 PgC in deep sediment accumulations below 3 m (Strauss et al., 2017).

This improved inventory 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 billion tons of additional carbon (Schuur et al., 2015). An additional pool is organic carbon remaining in permafrost but that is now submerged on shallow Arctic sea shelves that were formerly exposed as terrestrial ecosystems during the Last Glacial Maximum ~20,000 years ago (Walter et al., 2007). This permafrost is slowly degrading due to seawater intrusion, and it is not clear what amounts of permafrost and organic carbon still remain in the sediment versus what has already been converted to greenhouse gases. A recent synthesis of permafrost extent for the Beaufort Sea shelf showed that most remaining subsea permafrost in that region exists near shore with much reduced area (*high confidence*) as compared to original subsea permafrost maps that outlined the entire $3 x 10^6$ km^2 shelf area (<125 m depth) that was formerly exposed as land (Ruppel et al., 2016). These observations are supported by modelling that suggests that submarine permafrost would be already thawed >10 m depth or more under current submerged conditions (Anisimov et al., 2012; AMAP, 2017b).

3.4.1.3.4 Drivers of change in permafrost ground

Changes in temperature and precipitation act as gradual 'press' (i.e., continuous) disturbances that directly affect permafrost by modifying the ground thermal regime. Trends in temperature and precipitation in polar regions have been discussed earlier in the chapter, and the effects of those changes on permafrost are, for the most part, recorded in the observations of permafrost borehole temperatures (Biskaborn et al., 2015). Climate changes also can modify the occurrence and magnitude of abrupt physical disturbances such as fire, and soil subsidence and erosion resulting from ice-rich permafrost thaw. These 'pulse' (i.e., discrete) disturbances often are part of the ongoing disturbance and successional cycle in Arctic and boreal ecosystems (LTER, 2007), but changing rates of occurrence alter the landscape distribution of successional ecosystem states, with characteristic permafrost defined by the ecosystem and climate state (Jorgenson, 2013).

Pulse disturbances often rapidly remove the insulating soil organic layer, leading to permafrost degradation and loss in locations where permafrost temperature is already just below zero. Of all pulse disturbance types, wildfire affects the most high-latitude land area annually at the regional to 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). Fire activity is intimately coupled to climatic variation in regions where fuel buildup is not limiting to burning (van Leeuwen et al., 2014). There is *high confidence* that recent climate warming has been linked to increased wildfire activity in the boreal forest regions in Alaska (Kelly et al., 2013) and western Canadian (Kasischke and Turetsky, 2006; Flannigan et al., 2009) where this has been studied. Based on satellite imagery, an estimated 80,000 km^2 of boreal area was burned globally per year from 1997 to 2011 (van der Werf et al., 2010; Giglio et al., 2013). Extreme fire years in northern Canada during 2014 and Alaska during 2015 doubled the long-term (1997-2011) average area burned annually in this region, surpassing Eurasia to contribute 60% of the global boreal area burned (van der Werf et al., 2010; Mu et al., 2011; Randerson et al., 2012; Giglio et al., 2013). These extreme North American fire years were balanced by lower-than-average area burned in Eurasian forests, resulting in a 5% overall increase in global boreal area burned. There is *very high confidence* that changes in the fire regime are degrading permafrost faster than had occurred over the historic successional cycle (Rupp et al., 2016), and that the effect of this driver of permafrost change is underrepresented in the permafrost temperature observation network.

Abrupt permafrost thaw occurs when warming melts ground ice, causing the land surface to collapse. Ground subsidence alters surface hydrology; pooling or flowing water in turn causes more localized thawing and even mass erosion. Due to these localized feedbacks that can thaw through meters of permafrost within only a few years, permafrost thaw occurs much more rapidly than would be predicted from changes in air temperature alone. This is a pulse disturbance to permafrost that can occur in response to climate, such as an extreme precipitation event (Balser et al., 2014), or coupled with other disturbances such as wildfire that affects the ground thermal regime (Jones et al., 2015a). There is *low confidence* in the importance of abrupt thaw for driving change in permafrost ground at the circumpolar scale because it occurs at point locations rather than continuously across the landscape, but the risk for widespread change from this mechanism remains high because of the rapidity of change in these locations. New research at the global scale has revealed that 3.6x106 km^2, about 20% of the northern permafrost zone, appears to be vulnerable to abrupt thaw (Olefeldt et al., 2016). Abrupt thaw landforms are characteristic for wetlands, lakes, and hillslopes, where distinct thaw types form as a result of regional terrain and geomorphology (Kokelj et al., 2017; Shelef et al., 2017). Of the total area susceptible to abrupt thaw, wetland landscapes comprised 40%, lake landscapes 36%, and hillslope landscapes 25%. While 20% of the total permafrost zone was considered susceptible to abrupt thaw, the susceptible area contained 31% of the total organic carbon pool stored in the 0–3m soil and up to 50% of the the total carbon pool that includes the deep carbon >3 m, highlighting spatial correlation between processes and features that lead to abrupt thaw and storage of organic carbon.

3.4.2 Projected Changes to the Terrestrial Cryosphere

3.4.2.1 Seasonal Snow

Reductions in Arctic SCD are projected by the CMIP5 multi-model ensemble due to later snow onset in the fall and earlier snow melt in spring (Brown et al., 2017a) because of a warmer climate over essentially all Arctic land areas (Hartmann et al., 2013). There is *very high confidence* that projected declines are proportional to the amount of future warming in each model (Thackeray et al., 2016; Mudryk et al., 2017a). Projections to mid-century are primarily dependent on natural variability and model dependent uncertainties rather than the choice of forcing scenario (Hodson et al., 2013). By end of century, however, differences between scenarios emerge. RCP4.5 stabilizes at 5–10% Arctic SCD reductions (compared to a 1986–2005 reference period) while under RCP8.5, SCD continues to decline, reaching a –15 to –25% reduction by end of century (Brown et al., 2017a) (*very high confidence*). Within large initial condition ensemble experiments, individual realizations contain regions and seasons with cooling trends (Mudryk et al., 2014). Regions with negligible or increased SCD are therefore possible due to climate variability competing with anthropogenic forcing at the decadal and multi-decadal time scale.

There is *high confidence* CMIP5 models underestimate historically observed spring SCE reductions, due to uncertainty in the parameterization of snow processes (Essery, 2013; Thackeray et al., 2014) challenges in simulating snow-albedo feedback (Qu and Hall, 2014; Fletcher et al., 2015; Li et al., 2016), unrealistic temperature sensitivity (Brutel-Vuilmet et al., 2013; Mudryk et al., 2017a), and biases in climatological spring snow cover (Thackeray et al., 2016). The role of precipitation biases in influencing SCE projections is less well determined (Thackeray et al., 2016) (*medium confidence*).

Positive Arctic seasonal maximum pre-melt SWE (SWEmax) changes emerge across the eastern Eurasian Arctic by mid-century for both RCP4.5 and 8.5 (Brown et al., 2017a) (*high confidence*). Projected SWEmax increases across high latitude land areas of North America are less extensive and emerge later in the century, and only under RCP8.5 (Brown et al., 2017a). These projected SWEmax increases are due to enhanced snowfall (Krasting et al., 2013) from a more moisture rich Arctic atmosphere coupled with temperatures between November and April that remain sufficiently low for precipitation to fall as snow. This is not the case for May through October, and for more temperate regions of the Arctic (i.e., Scandinavia) where temperatures do not remain sufficiently low and precipitation phase changes to rainfall result in projected decreases in SWE (de Vries et al., 2014; Brown et al., 2017a). SWE across large portions of the Arctic is presently unaffected by temperature variability (SWE is therefore solely driven by precipitation availability) but this area is projected to decrease by mid-century as temperature forcing of precipitation phase becomes more important across larger regions of the Arctic and for longer periods of the shoulder seasons (Sospedra-Alfonso and Merryfield, 2017). Projected increases in high latitude snowmelt runoff are *very likely* to result from increased SWE (Mankin et al., 2015).

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

3.4.2.2 Freshwater Systems

Climate model simulations project a warmer and wetter Arctic (Krasting et al., 2013). Specific humidity is projected to increase, as warming temperatures in the lower troposphere drive enhanced evaporation (Laîné et al., 2014), and moisture flux convergence increases into the Arctic (Skific and Francis, 2013) (*high confidence*). Relative humidity changes will be driven by contrasts in heating over land versus ocean, and the influence of this heating on marine air masses advected over land. These processes are poorly resolved in climate models, so there is only *medium confidence* in relative humidity projections (Vihma et al., 2016).

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

Changes in lake ice are not well represented in global climate models because individual lakes are not resolved. When forced with regional climate models, lake ice models project an earlier spring break-up of between 10–25 days by mid-century (compared with 1961–1990), and up to a 15-day delay in the freeze-up for lakes in the North American Arctic (Brown and Duguay, 2011; Dibike et al., 2011; Prowse et al., 2011b). This results in a reduction of ice cover duration of 10–35 days under RCP4.5. More extreme reductions of up to 60 days are projected in coastal regions. Mean maximum ice thickness is projected to decrease by 10–50 cm over the same period (Brown and Duguay, 2011). High-latitude warming is projected to drive earlier river ice break-up in spring due to both decreasing ice strength, and earlier onset of peak discharge (Cooley and Pavelsky, 2016). The complex interplay between hydrology and hydraulics in controlling spring flooding and ice jam events (which can be important events for sediment and nutrient transport; Turcotte et al. (2011)) reduce confidence in related projections (Prowse et al., 2010; Prowse et al., 2011b).

3.4.2.3 Permafrost and fire

Models at the circumpolar or global scale represent permafrost degradation in response to warming as increases in active layer thickness. As projected air temperatures continue to rise, the active layer becomes too deep to completely refreeze during winter, forming a layer called a talik (Sazonova et al., 2004; Schaefer et al., 2011). The CMIP5 models project with *high confidence* that active layers will increase and areal extent of near-surface permafrost will decrease substantially (Lawrence et al., 2012; Koven et al., 2013). However, there is only *medium confidence* in the magnitude of these changes due to a four-fold range of estimated present day permafrost area by these models using the same climate scenario. This was caused by wide range of model sensitivity in permafrost area to air temperature change, resulting in a large range of projected permafrost loss: 15–87% under RCP4.5 and 30–99% under RCP8.5 (Slater and Lawrence, 2013). estimated that the reduction in near-surface permafrost extent by the end of the century will be about 2.1 million km^2 under RCP 2.6 and 10 million km^2 under RCP 8.5. The high warming scenario would leave most of the current discontinuous permafrost zone free of near-surface permafrost with the remaining permafrost located around around the coldest regions in the northern hemisphere: Northern Siberia and the islands of Northeast Canada. A more recent analysis of permafrost trends from a subset of models that self-identified as structurally representing the permafrost zone had a significantly smaller range of estimated present day permafrost area (13.1-19.3 million km^2)(McGuire et al., 2016). This subset also showed large reductions of permafrost area under RCP8.5, averaging a loss of 12.7 million km^2 of permafrost area by 2300, with much of that long-term loss already occurring by 2100 (McGuire et al., 2018).

Pulse disturbances are not included in the permafrost projections described above, and there is *high confidence* that fire and abrupt thaw will accelerate change in permafrost ground relative to climate effects alone, if the rates of these disturbances increased. Large interannual variability in the fire regime makes long-term trends difficult to identify in the fire record, but there is *high confidence* that fire frequency and area burned have increased in recent years in Alaska and Western Canada, with new regional records for area burned set in 2004, 2014, and 2015 (Rupp et al., 2016; Walker et al., in press). This trend is projected to continue for the rest of the century across most of the tundra and boreal region for many climate scenarios, with the boreal region projected to have the greatest increase in total area burned (Balshi et al., 2009; Rupp et al., 2016). Due to vegetation-climate interactions, there is only *medium confidence* in projections of future area burned. As fire activity increases, flammable vegetation, such as the black spruce forest that dominates boreal Alaska, is projected to decline as it is replaced by low-flammability deciduous forest (Johnstone et al., 2011). In other regions such as western Canada, by contrast, black spruce could be replaced by the even more flammable jack pine, creating regional-scale feedbacks that increases the spread of fire on the landscape. In tundra regions, graminoid tundra is projected to be replaced by more-flammable shrub tundra in future climate scenarios, and tree migration into tundra could further increase fuel loading (Rupp et al., 2016). In contrast to fire, there are no regional or global projections of how abrupt thaw rates may change in the future and so there is *low confidence* in the ability to assess risk, even though this mechanism for rapid change appears critically important for projecting future change (Kokelj et al., 2017).

3.4.3 Consequences of Change to the Terrestrial Cryosphere

3.4.3.1 Global Climate Feedbacks

3.4.3.1.1 Changing carbon cycle: carbon dioxide and methane

Climate warming is expected to change the storage of carbon in vegetation and soils in northern regions, and net carbon transferred to the atmosphere as carbon dioxide or methane acts a feedback to accelerate global climate change. There is *high confidence* that the northern region historically acted as net carbon sink as carbon accumulated in terrestrial ecosystems since the Last Glacial Maximum (Schirrmeister et al., 2011; Hugelius et al., 2014; Loisel et al., 2014). There is increasing, but divergent evidence, that changing climate in the modern period has shifted these ecosystem into net carbon sources (*low confidence*). Syntheses of ecosystem CO_2 fluxes have alternately showed tundra ecosystems as carbon sinks or neutral averaged across the circumpolar region for the 1990s and 2000s (McGuire et al., 2012), or carbon sources over the same time period (Belshe et al. 2013). Both syntheses agree that the summer growing season is a period of net carbon uptake into terrestrial ecosystems (*high confidence*), and this uptake appears to be increasing (Belshe et al., 2013; Ueyama et al., 2013). The discrepancy may be a result of CO_2 release rates 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 ecosystems types that can differ in carbon sink/source strength. Recent 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 forests were either neutral or a net CO_2 sink for the period 2012 to 2014 (Commane et al., 2017). The Alaska study region as a whole was estimated to be a net carbon source of 25 ± 14 Tg C per year averaged over the land area of both biomes for the entire study period. If this Alaskan region (1.6 x 10^6 km^2) was representative of the entire northern circumpolar permafrost zone soil area (17.8 x 10^6 km^2), this would be equivalent to a region-wide net source of 0.3 Pg C per year.

The permafrost soil carbon pool is climate sensitive and an order of magnitude larger than carbon stored in plant biomass (*high confidence*)(Schuur et al. 2018 State of Carbon Cycle Report, in revision). Initial estimates were converging on an range of cumulative emissions from soils to the atmosphere, but recent studies have actually widened that range somewhat (*medium confidence*) (Figure 3.14). Expert assessment and lab incubation studies suggest that substantial quantities (tens to hundreds Pg) of C could potentially be transferred from the permafrost carbon pool into the atmosphere under a warming climate (RCP8.5)(Schädel et al., 2014; Schädel et al., 2016; Schuur et al., in review). Global dynamical models supported these findings, showing potential carbon release from the permafrost zone ranging from 37 to 174 PgC by 2100 under the current climate warming trajectory (RCP8.5), with an average across models of 92 ± 17 PgC (mean ± SE) (Zhuang et al., 2006; Koven et al., 2011; Schaefer et al., 2011; Burke et al., 2012; MacDougall et al., 2012; Burke et al., 2013; Schaphoff et al., 2013; Schneider von Deimling et al., 2015). This range is generally consistent with several newer data-driven modeling approaches that estimated that soil carbon releases by 2100 (for RCP8.5) will be 57 PgC (Koven et al., 2015) and 87 PgC (Schneider von Deimling et al., 2015), as well as an updated estimate of 102 PgC 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., 2017), or common environmental input data (McGuire et al., 2016) show similar soil carbon losses, but also indicate the potential for stimulated plant growth (nutrients, temperature/growing season length, CO_2 fertilization) to offset some or all of these losses by sequestering new carbon into plant biomass and increasing inputs into the surface soil (McGuire et al., 2018). Overall, the estimates support the idea that the northern permafrost zone could emit carbon on the order similar to other current biospheric sources like land use change, but will generally be only a fraction of fossil fuel emissions (*high confidence*). Furthermore, there is *high confidence* that climate scenarios that involve mitigation (e.g RCP4.5) will help to slow the emissions of carbon from the Arctic and boreal regions (Schuur et al., 2015; Schuur et al., in review).

There is *low confidence* about the degree to which additional CH_4 from northern lakes, ponds, and wetland ecosystems is already contributing to increasing atmospheric concentrations. Long-term direct observations of CH_4 dynamics are scarce, and analyses of atmospheric concentrations in Alaska concluded that local ecosystems surrounding the observation site have not changed in the exchange of CH_4 from the 1980s until the present, which suggests that either the local wetland ecosystems are responding in step with other northern wetland ecosystems, or that increasing atmospheric CH_4 concentrations in northern observation sites is derived from CH_4 coming from mid latitudes (Sweeney et al., 2016). In contrast, indirect integrated estimates of CH_4 emissions from expanding permafrost thaw lakes suggest a release of an additional 1.6–5 Tg CH_4 per year over the last 60 years (Walter Anthony et al., 2014). At the same time, there is *high confidence* that we have been under-observing CH_4 reflected in new quantifications of cold-season methane emissions can be >50% of the annual budget (Zona et al., 2016), as well as of geological CH_4 seeps that may or may not be climate sensitive (Walter Anthony et al., 2012; Kohnert et al., 2017). Observations such as these underlie the fact that source estimates for CH_4 made from atmospheric observations do not match CH_4 source estimates made from upscaling of ground observations at the global scale, and this problem has not improved over several decades of research (Crill and Thornton, 2017).

In many of the model projections previously discussed, CH_4 release is not explicitly represented because fluxes are small even though higher GWP of CH_4 makes these emissions relatively more important than on a mass basis alone. Global models that include the short-term sensitivities of CH_4 to warming show increased CH_4 emissions to future warming in the northern permafrost region (Riley et al., 2011; Gao et al., 2013). Yet, these models conclude that if these increased emissions were to occur, they would have little influence on the climate system because of their relatively small magnitude. However, most models do not include abrupt thaw processes that can result in lake expansion, wetland formation, and massive erosion and exposure to decomposition of previously frozen carbon-rich permafrost, leading to *low confidence* in future model projections of CH_4. A substantial area of the northern permafrost region is susceptible to abrupt thaw (Olefeldt et al., 2016), which could result in more substantial CH_4 emissions in the future than are currently projected by models. A recent study that does include these processes suggests that the largest CH_4 emission rates will occur around the middle of this century when simulated thermokarst lake extent is at its maximum and when abrupt thaw under thermokarst lakes is taken into account (Schneider von Deimling et al., 2015). Furthermore, the simulated CH_4 fluxes in can cause up to 40% of total permafrost-affected radiative forcing in this century. Similarly, no global models currently consider the effects of warming on CH_4 emissions from coastal and ocean shelf systems in the Arctic.

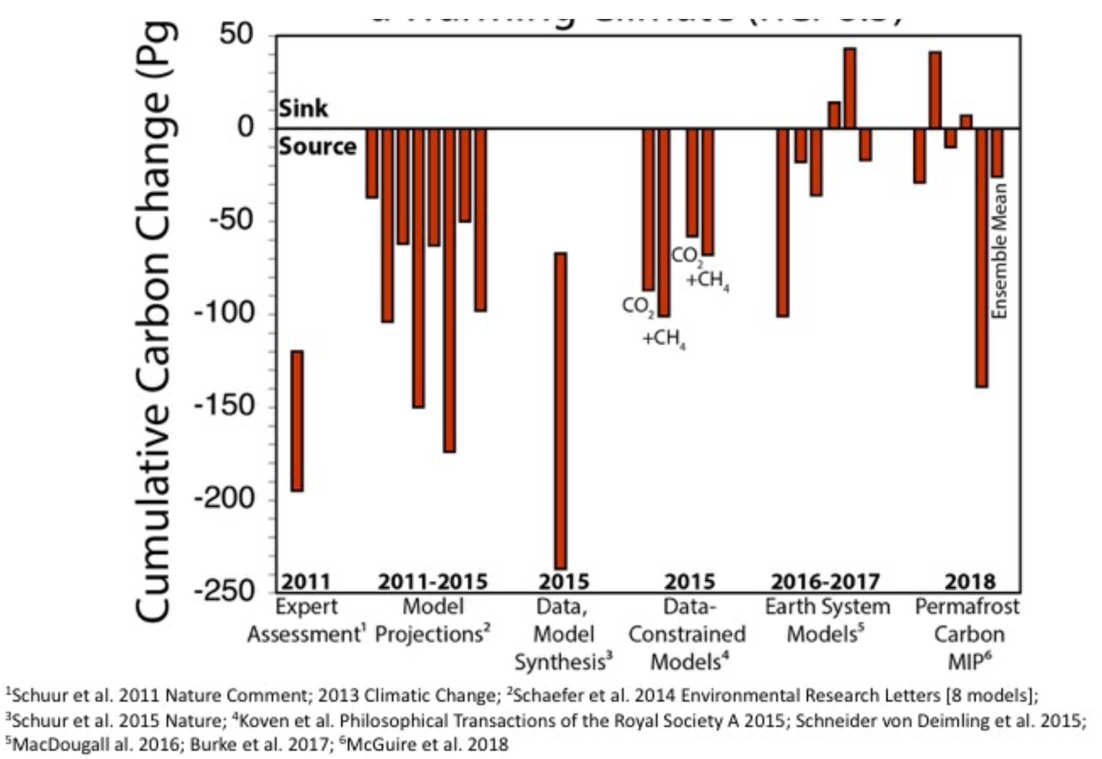


Figure 3.14. Estimates of cumulative net soil carbon pool change for the northern circumpolar permafrost zone by 2100 following the RCP8.5 warming scenario. Cumulative carbon amounts are shown in Petagrams C (1 Pg C=1 billion metric tons), with **source** (negative values) indicating net carbon movement from soil to the atmosphere and **sink** (positive values) indicating the reverse. Data are from [1]Schuur et al. 2011 Nature Comment; Schuur et al. (2013); [2]Schaefer et al. (2014) [8 models]; [3]Schuur et al. (2015); [4]Koven et al. (2015); Schneider von Deimling et al. (2015); [5]MacDougall and Knutti (2016); Burke et al. (2017); [6]McGuire et al. (2018)

3.4.3.1.2 Changing energy budget

Warming induced reductions in the duration and extent of spring snow cover (Section 3.4.1.1) lowers albedo because snow-free land reflects much less solar radiation than snow. The corresponding increase in net radiation at the surface constitutes a small positive feedback to global climate associated with an increase in atmospheric sensible heating (Flanner et al., 2011; Qu and Hall, 2014; Thackeray and Fletcher, 2016). The influence of snow albedo on the planetary global energy budget can be quantified using the snow shortwave radiative effect (SSRE). Over 1979-2008, changes in snow cover led to an increase in global net solar energy flux at both the surface and top of atmosphere (TOA; *high confidence*) estimated to be 0.22 W m^{-2} (±50%; *medium confidence*) which weakened the hemispheric TOA SSRE, (Flanner et al., 2011). Since AR5, Chen et al. (2015) calculated a 2001−2013 weakening in NH SSRE of 0.16 W m^{-2}, (similar to an estimate of 0.10 W m^{-2} during 1982−2013 by Chen et al. (2016)), while Singh et al. (2015) did not find a statistically significant trend in global SSRE (2001−2013). A key source of uncertainty in these SSRE change calculations is the range in observed spring snow cover extent trend estimates (Hori et al., 2017; Mudryk et al., 2017a).

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

A notable anthropogenic forcing mechanism operating directly on the climate system is the deposition of black carbon (BC) and other light absorbing impurities in seasonal snow. Subsequent darkening of snow causes it to melt earlier and drive positive albedo feedback (e.g., Hansen and Nazarenko, 2004). Based largely on the industrial-era (1750−2010) radiative forcing of +0.035 W m^{-2} derived by Bond et al. (2013) for BC in land-based snow, the IPCC AR5 adopted a global direct radiative forcing of 0.04 W m^{-2} for BC in snow and sea-ice, but noted that the effective forcing is about 3-fold greater than the direct forcing due to a strong albedo feedback triggered by the initial darkening. Lawrence et al. (2011) estimate the present-day radiative effect of BC and dust in land-based snow to be 0.083 W m^{-2}, only marginally greater than the simulated 1850 effect (0.075 W m^{-2}) due to offsetting effects from increased BC emissions and reductions in dust darkening and snow cover (*medium confidence*). Lin et al. (2014) provide the first-ever estimate of forcing from brown carbon deposited in snow (associated with both combustion and secondary organic carbon) finding a range of 0.9−2.5 m W m^{-2} associated with different assumptions of particle absorptivity (*low confidence*).

3.4.3.2 Ecosystems and their Services

3.4.3.2.1 Aquatic ecosystems

Changes in the terrestrial cryosphere have direct impacts on Arctic aquatic ecosystems and associated services. Changes in aquatic greening and browning have important implications for water quality and food provisioning in northern communities. Increases in riparian vegetation (birch, willow, alder) along Arctic river corridors (‘shrubification’ (Myers-Smith et al., 2015)) is expected to enhance inputs of terrestrial allochthonous carbon to stream networks (Wrona et al., 2016). This enhanced nutrient input from riparian shrubs is expected to stimulate lotic food webs, increasing the productivity of microbial decomposers and invertebrate detritivores (Wrona et al., 2016). Furthermore, expansion of nitrogen-fixing alders along riparian corridors could result in additional nitrogen input into riparian and aquatic systems (Tape et al., 2006).

The role of changing water sources with respect to land ice, snow melt, and groundwater will influence biological communities (Blaen et al., 2014a). In the Arctic snow melt dominated streams, the size of the winter snowpack can influence benthic communities and cause significant inter-stream differences in the same year and intra-stream differences from year to year (Docherty et al., 2017). There is *high confidence* that glacier-fed rivers are presently experiencing sustained (though finite) periods of increased discharge

(Liljedahl et al., 2016) leading to more favorable habitats for some invertebrate and fish species (Vincent et al., 2011). Projected increases in baseflow resulting from permafrost thaw and consequent reduced runoff to infiltration ratios, are likely to have a similar effect in sustaining seasonal flow and regulating stream temperatures (Walvoord and Kurylyk, 2016) (*high confidence*). If conditions become less harsh in streams of the Arctic, potentially more species could be supported but dispersal constraints related to biogeography limit potential colonization (Hotaling et al., 2017). In regions with pronounced organic layers, increases in surface water connectivity, attributed to peat plateau collapse from permafrost thaw (Connon et al., 2014) will enhance the through-flow of nutrients (*high confidence*). Enhanced transfer of nutrients from land to water (driven by active layer thickening and thermokarst processes; Abbott et al. (2015); Vonk et al. (2015)) is leading to heightened autotrophic productivity in freshwater ecosystems (Wrona et al., 2016). There is *low confidence* on the influence of permafrost changes on DOC. Permafrost thawing and increased depth of the active layer could enhance transmission of DOC to stream systems which could facilitate ammonium retention in these systems resulting in less export to the ocean with climate change (Blaen et al., 2014b). Conversely, reduced DOC export could accompany permafrost thaw as (1) water infiltrates deeper in the subsurface and has longer residences times for DOC decomposition (Striegl et al., 2005) and (2) the proportion of groundwater (typically lower in DOC, higher in DIC than near surface runoff) to total streamflow increases (Walvoord and Striegl, 2007). Emerging evidence suggests large stores of mercury in permafrost may be released upon thaw, thereby having effects (largely unknown at this point) on aquatic ecosystems (Schuster et al., 2018).

Legacy pollutants like black carbon, POPs (e.g., HCHs, PAHs, PCBs, etc.) can be transferred downstream and affect water quality (Hodson, 2014). Lakes can become sinks of these contaminants, while important floodplains can be contaminated (Sharma et al., 2015). In lentic systems, an extended growing season for plankton and macrophytes affects water quality and aquatic community structure of inland waters. Shortened duration of snow and ice cover (more light absorption, increased nutrient input) is expected to result in higher primary productivity (Hodgson and Smol, 2008; Vincent et al., 2011). Shifts in the surface water balance are also being observed – permafrost thaw is resulting in drying/draining of lakes and wetlands in some areas, and elsewhere is contributing to the creation of thaw collapse lakes and wetlands (see Section 3.4.1.2.2).

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

Habitat loss or change due to climate warming are serious threats to Arctic fishes. Surface water loss, reduced surface water connectivity among aquatic habitats (streams, lakes, ponds, wetlands), and changes to the timing and magnitude of seasonal flows (see Section 3.4.1.2) results in a direct loss of spawning, feeding, or rearing habitats, or access by fishes to them (Poesch et al., 2016). Changes to permafrost-dominated landscapes, including the transition from surface water-dominated systems to ground water-dominated systems in some regions (Frey and McClelland, 2008) has reduced freshwater habitats available for fishes and other aquatic biota, including the aquatic invertebrates upon which the fish depend for food. Gullying deepens channels (Rowland et al., 2011a; Liljedahl et al., 2016) that otherwise may connect lentic habitats occupied by fishes, which in turn can lead to loss of surface water connectivity and the inability for fish to potentially access key habitats. Surface water connectivity controls fish species occurrence and assemblages in lentic systems across the Arctic landscape (Haynes et al., 2014; Laske et al., 2016), and less habitat connectivity means lower fish diversity. These small connecting channels, which are more vulnerable to drying than larger ones, 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 seasonal hydrological events, especially high flows that play important roles triggering and allowing fish dispersal and migrations is also a threat. Arctic fish dispersal and migration activities are timed around high surface flow events in early and late summer (Heim et al., 2016). If the timing of an important life history event such as spawning becomes mismatched with changing stream flows, they will likely be negatively affected if they cannot adapt to seasonally shifted flow regimes. Changes to the Arctic growing season (Xu et al., 2013a) increases the risk of drying of surface water habitats and poses a potential mismatch in seasonal availability of food in rearing habitats.

Lentic and lotic systems across the Arctic are by nature relatively shallow, and are expected to warm further with longer growing seasons. This may make some of the 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; invertebrate life histories and production are temperature and degree-day dependent, although there are no reports on how food resources have been affected, and making reliable projections extremely difficult. Increased nutrient export from permafrost loss (Frey et al., 2007), facilitated by warmer temperatures, will likely increase food resources for consumers, but how that impacts lower trophic levels within food webs remains speculative. While long-term data on run timing of fishes are limited, phenological shifts could result in mismatches with food availability or habitat suitability in both marine and freshwater environments for anadromous species, and in freshwater environments for freshwater-resident species.

3.4.3.2.2 Terrestrial ecosystems

Vegetation

Changes in tundra vegetation can have important effects on permafrost, hydrological dynamics, carbon and nutrient cycling, and the surface energy balance (e.g., Myers-Smith and Hik, 2013; Frost and Epstein, 2014), as well as the diversity, abundance, and distribution of both wild and domesticated herbivores (e.g., Fauchald et al., 2017; Horstkotte et al., 2017) in the Arctic. There is *high confidence* that the overall trend for tundra vegetation in the 35-year satellite record (1982-2016) shows increasing aboveground biomass (=greening) throughout a majority of the geographic circumpolar arctic tundra (Xu et al., 2013a; Ju and Masek, 2016; Bhatt et al., 2017). Regions with some of the greatest increases in tundra greenness are the North Slope of Alaska, the Low Arctic (southern tundra subzones) of the Canadian tundra, and east of the Taimyr Peninsula in north-central Siberia, Russia. Increasing greenness observed by satellites has been linked with *high confidence* to changes in plant community composition shifting dominance away from graminoids towards shrubs (Myers-Smith et al., 2015). Despite the overall trend of greening, there are regions where tundra aboveground vegetation biomass has declined (=browning) including on the Yukon-Kuskokwim Delta of western Alaska, the High Arctic of the Canadian Archipelago, and the northwestern Siberian tundra (Bhatt et al., 2017).

Greening and browing trends in tundra vary spatially and are not always consistent over the decadal scales of the observational record, suggesting complex interactions among atmosphere, ground (soils and permafrost), vegetation, and herbivores that control these responses. There is *high confidence* that controls over tundra greening include increases in summer, spring, and winter temperatures, as well as growing season length (e.g., Vickers et al., 2016; Myers-Smith and Hik, 2018), in part linked to reductions in Arctic Ocean sea-ice cover (Bhatt et al., 2017; Macias-Fauria et al., 2017). Other controls on tundra greening include increases in snow water equivalent and soil moisture, increases in active layer depth (nutrient availability), changes in herbivore activity, and to a lesser degree human use of the land (e.g., Salmon et al., 2016; Horstkotte et al., 2017; Martin et al., 2017; Yu et al., 2017). Changes in the phenology of tundra vegetation are also apparent, and these are commonly related to changes in snow cover and the timing of snowmelt, although these also tend to vary with other environmental factors (Oberbauer et al., 2013; Bhatt et al., 2017; Prevéy et al., 2017). Research on tundra browning is more limited but suggests that changes in winter climate – specifically reductions in snow cover due to winter warming events that expose tundra to subsequent freezing and desiccation, in addition to disturbances such as insect and pathogen outbreaks, increased grazing pressure, and surface and ground ice thawing that increases surface water (Phoenix and Bjerke, 2016; Bjerke et al., 2017)(*medium confidence*).

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

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

Wildlife

Wild reindeer and caribou (Rangifer), 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. The culture and subsistence of indigenous Arctic people co-evolved with the Rangifer seasonal migrations. Migratory tundra wild reindeer/caribou have declined from about 5 million in the 1990s to the current status (2017) of about 2 million on the continental USA (Alaska), Canada, Greenland and Russia (https://carma.caff.is/herds). Published population estimates and trends exist for some, but not all herds (Pachkowski et al., 2013; Adamczewski et al., 2015; Nicholson et al., 2016; Tyson, 2016). For the Arctic Islands, the current estimate is approximately 56,000, which represents a long-term decline since the 1960s and 70s. Within the overall decline, numbers have recently increased on some High Arctic Islands (including Svalbard); the Porcupine Caribou herd, straddling Yukon and Alaska, is at a historic high (https://carma.caff.is/herds). While wild reindeer and caribou abundance cycle naturally over 40-60 year periods, the current rates of decline and low numbers exceed historical declines for many herds (*high confidence*).

The drivers of the observed Rangifer decline are complex, often with time lags and carryover effects. Climate strongly influences productivity and in some years, extremes in heat, drought, winter icing, and/or deep snow reduce survival (Mallory and Boyce, 2017). In particular, snow layering (particulary ice lenses from rain on snow events; compacted snow from heavy snowfall events) can negatively impact forage and mobility (Riseth et al., 2011; Forbes et al., 2016). Rain-on-snow (ROS) events, with resulting ice-encrusted rangelands, can lead to catastrophic mass starvation of Rangifer (Forbes et al., 2016; Bartsch et al., 2017). Late and weak ice formation on waterbodies can impact rangifer herding activities (Turunen et al., 2016). At the same time, summer warming is increasing both plant growth and changing the composition of tundra plant communities, modifying the relationship between climate, forage and Rangifer (Albon et al., 2017). These changes then propagate through the ecosystem with effects on other herbivores such as geese, voles, and predators (Hansen et al., 2013).

Changes in the timing of sea-ice formation have direct effects on Rangifer migration and survival. Sea ice seasonally links the Arctic Basin's six archipelagos and coastal islands (totaling 1 million km^2). Rangifer depend on sea-ice for their inter-island movements and migrations to the continental mainland. For example, sea ice now forms 8–10 days later than it did in the early 1980's between Victoria Island and the mainland, so caribou of the Dolphin and Union herd now cross the strait closer to the time of initial ice formation, which increases risks of breaking through the ice (Poole et al., 2010).

In northern Fennoscandia, the reindeer population totals around 600,000 animals, managed by indigenous Sami (Norway, Sweden) and a mix of Sami and non-indigenous herders (Finland). Lichen rangelands lie at the nexus of debate over reindeer "carrying capacity" in both Fennoscandia and northern Russia. There is *low confidence* in lichen response to climate change: measurements suggest enhanced summer precipitation increases lichen biomass, while an increases in winter precipitation lowers it (Kumpula et al., 2014).

The observed decline of Rangifer has profound implications for the food security and cultures of indigenous people (Horstkotte et al., 2017; Lavrillier and Gabyshev, 2017). While the resilience of Rangifer to changing landscapes (amid an intensifying human footprint across the Arctic) has been reduced, effective management can increase resilience through building adaptive capacity such as maintaining the connectivity of seasonal ranges and implementing trade-offs between factors affecting Rangifer. In semi-domesticated populations, management strategies may have masked the effects of climate (Uboni et al., 2016). A recent review of Russian reindeer populations reached a similar conclusion, socio-economic factors conceal the impact of climate change on reindeer populations (Klokov, 2012).

3.4.3.3 Sectoral Consequences of a Changing Terrestrial Cryosphere

The Circumpolar Arctic is home to over four million people, with large regional variation in population distribution and demographics (Heleniak and T., 2014). 'Connection with nature' is a defining feature of Arctic identity (Schweitzer et al., 2014), particularly amongst Indigenous populations for which the land, ocean and cryosphere which surround communities evoke a sense of home, freedom, and belonging (Cunsolo Willox et al., 2012; Cunsolo Willox et al., 2013; Durkalec et al., 2015). There is very high confidence that climate-driven environmental changes that affect local ecosystems influence travelling, hunting, fishing, and gathering, with implications for people's lives, lifestyles, cultural practices, economies, and self-determination.

3.4.3.3.1 Subsistence harvesting, food and water security

Impacts of climate change on food and water security in the Arctic are more severe in regions where infrastructure, 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). Impacts are highly variable based on local community and cultural contexts, as well as regional geographic differences in environmental and climatic conditions.

Food security

There is *high confidence* that food insecurity is on the rise for Arctic peoples (Council of Canadian Academies, 2014; Rautio et al., 2014). Climate change consequences for northern ecosystems combined with processes of globalization and complex social, economic and cultural factors contribute to a dietary transformation from locally resourced foods to imported market foods across northern regions in recent decades (Harder and Wenzel, 2012; Parlee and Furgal, 2012; Nymand and Fondahl, 2014; Beaumier et al., 2015). Food systems of northern communities are intertwined with northern ecosystems because of traditional and subsistence hunting, fishing, and gathering activities. Environmental changes to animal habitat and movement mean that important food species may no longer be found within accessible ranges or familiar areas (Parlee and Furgal, 2012; Rautio et al., 2014; Inuit Circumpolar Council, 2015; Overland et al., 2017a) (Section 3.4.3.2.2). This impacts the accessibility of culturally important local food sources (Rosol et al., 2016) that make important contributions to a nutritious diet (Donaldson et al., 2010; Parlee and Furgal, 2012; Hansen et al., 2013; Dudley et al., 2015). Rain on snow events are a particular challenge for caribou and reindeer to access forage (Hansen et al., 2014; Overland et al., 2017a; Overland et al., 2017b) with impacts on animal health, mortality, and meat quality in commercial reindeer herding operations (Hansen et al., 2014). Enhanced vegetation growth and expansion into more northern latitudes (Section

3.4.3.2.2) create more food for animals such as moose, but impacts other plants such as lichens upon which caribou depend (Parlee and Furgal, 2012). Longer open water seasons and poorer ice conditions on lakes (Section 3.4.1.2) impact fishing options (Laidler, 2012) and waterfowl hunting (Goldhar et al., 2014).

There is *high confidence* that changes to travel conditions impact food security through access to hunting grounds. Shorter snow cover (Section 3.4.1.1), and changes to snow conditions (such as density), and earlier ice break-up (Section 3.4.1.2) make overland travel more difficult and dangerous (Ford and Pearce, 2012; Laidler, 2012; Parlee and Furgal, 2012; Cunsolo Willox et al., 2013; Ford et al., 2016; Overland et al., 2017a). Changes in dominant wind direction and unpredictable winds reduce the reliability of traditional navigational indicators such as snow drifts, increasing safety concerns (Ford and Pearce, 2012; Laidler, 2012; Ford et al., 2013; Ford et al., 2016; Clark et al., 2016b). Permafrost warming, increased active layer thickness (Section 3.4.1.3), and changes to water levels (Section 3.4.1.2) impact overland navigability in summer (Goldhar et al., 2014). Of particular concern for coastal communities is landfast sea ice (Section 3.3.1.1.5), including the floe edge position, timing and dynamics of freeze-up and break-up, stability through the winter, and then length of summer open water season (Gearheard et al., 2013; Eicken et al., 2014; Alaska, 2015a; Ford et al., 2016; Baztan et al., 2017). Rough surface conditions and the presence of ridges characterize "bad" ice that is difficult to travel on (Cunsolo Willox et al., 2013). Warming water temperature, altered salinity profiles, snow properties, changing currents and winds all have consequences the use of sea ice as a travel or hunting platform (Hansen et al., 2013; Eicken et al., 2014; Alaska, 2015a; Clark et al., 2016a).

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

Water security

Drinking water quantity and quality is a concern across the Arctic (Nymand and Fondahl, 2014) due to the challenges of water treatment, transport, maintenance, and supply to growing communities (Cochran et al., 2013; Goldhar et al., 2013; Cunsolo Willox et al., 2015; Daley et al., 2015; Dudley et al., 2015). Many northern communities rely on ponds, streams, and lakes for drinking water (Cochran et al., 2013; Goldhar et al., 2013; Goldhar et al., 2014; Alaska, 2015b; Daley et al., 2015; Overland et al., 2017b), so projected changes in seasonal precipitation and hydrology will impact water supply (*high confidence*) (Section 3.4.2.2). Surface water is vulnerable to thermokarst disturbance and drainage, as well as bacterial contamination, both of which could be impacted by warming ground and water temperatures (Cozzetto et al., 2013; Goldhar et al., 2013; Dudley et al., 2015; Overland et al., 2017b). Icebergs or old multi-year ice are important sources of drinking water for coastal communities, so the changing presence or accessibility to access these water resources also affect local water security. Vulnerable water security is amplified by the reduced capacity of small remote communities to respond quickly to water supply threats (Daley et al., 2015).

3.4.3.3.2 Communities

Culture and knowledge

Spending time on the land is culturally important for Indigenous communities (Eicken et al., 2014; Durkalec et al., 2015; Inuit Circumpolar Council, 2015). There is *very high confidence* that climate change impacts daily life because land-based activities and community events are closely tied to seasonal cycles connected to ice freeze-up and break-up (rivers/lakes/sea ice), snow onset/melt, vegetation stages, and related wildlife/fish/bird behaviour (Inuit Circumpolar Council, 2015). Inter-generational knowledge transmission of associated values and skills is also influenced by climate change (Ford and Pearce, 2012; Eicken et al., 2014; Cunsolo Willox et al., 2015; Inuit Circumpolar Council, 2015). Where changes are happening rapidly or unpredictably, younger generations do not have the same level of experience or confidence with traditional indicators (Ford, 2012; Parlee and Furgal, 2012; Cunsolo Willox et al., 2015; Ford et al., 2016). This erodes confidence in traditional knowledge and knowledge holders (Ford and Pearce, 2012; Parlee and Furgal, 2012; Cunsolo Willox et al., 2015; Ford et al., 2016) leading to emotional and cultural responses (Cunsolo Willox et al., 2015).

Economics

The northern mixed economy is characterized by a combination of subsistence activities and employment income (Cunsolo Willox et al., 2012; Ford and Pearce, 2012; Harder and Wenzel, 2012; Cochran et al., 2013; Fall, 2016; Ford et al., 2016; Clark et al., 2016b). The social economy related to sharing, kinship, and the framing of household economic conditions has received limited research attention (Ford et al., 2012; Harder and Wenzel, 2012; Fall, 2016). It is difficult to assess how climate change will impact on local subsistence activities and economic opportunities because of high level of differences between communities and because the impact of climate change on subsistence contributions to the mixed economy is not well understood.

Longer ice-free travel windows in Arctic seas could lower the costs of access and development of northern resources (delivering supplies and shipping resources to markets) and thus may contribute to increased opportunities for marine shipping, commercial fisheries, tourism, and resource development (Ford et al., 2012; Huskey et al., 2014; Ford et al., 2016; Overland et al., 2017b). This has important implications for environmental impacts and economic development, particularly in relation to local employment opportunities or concerns of detrimental impacts on animals, habitat, and subsistence activities (Cochran et al., 2013; Inuit Circumpolar Council, 2015). There are also many associated risks with unpredictable sea ice conditions, and development costs could remain high due to increased flooding, coastal erosion, and impacts on infrastructure (Huskey et al., 2014).

3.4.3.3.3 Health and well-being

For many polar residents, especially Indigenous peoples, the physical environment underpins social determinants of well being and physical and mental health. Changes to the environment impacts most dimensions of health and wellbeing (Driscoll et al., 2013; Parlee et al., 2012).

Injury and death

Climate change consequences for ice cover in polar regions (Section 3.3.1.1; 3.4.1.2) have impacted key transportation routes (Clark, et al., 2016; Ford et al., 2013; Gearheard et al., 2006; Laidler, 2006), and pose increased risk of injury and death during travel (Clark et al., 2017; Driscoll et al., 2016, 2013, Durkalec et al., 2014, 2015).

Foodborne disease

Non-infectious 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, foodborne gastroenteritis is associated with shellfish harvested from warming waters (McLaughlin et al., 2005; Young et al., 2015). Permafrost warming and increases in active layer thickness (Section 3.4.1.3) reduce the reliability of permafrost for natural refrigeration. This reduces access to and consumption of locally resourced food, and results in increased incidence of illness from spoiled meat (Laidler, 2012; Cochran et al., 2013; Cozzetto et al., 2013; Rautio et al., 2014; Beaumier et al., 2015; Overland et al., 2017a).

Waterborne disease

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) (Brubaker et al., 2011; Dudley et al., 2015; Parkinson et al., 2005, 2009). After periods of rapid snowmelt, bacteria can increase in untreated drinking water, with associated increases in acute gastrointestinal illness (Harper et al., 2011). Consumption of untreated drinking water may increase duration and frequency of exposure to local environmental contaminants or potential water- or vector-borne diseases (Goldhar et al., 2014; Daley et al., 2015). The potential for infectious gastrointestinal disease is not well understood, and there may be greater concerns in relation to storage containers of raw water than the source water itself (Goldhar et al., 2014).

Mental health and wellbeing

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). This reflects the concept of 'solastalgia' - the "pain, distress, and sadness that emerges when a place to which individuals and groups are deeply and intimately attached changes in a manner that disrupts opportunities for place-based solace, comfort, and familiar activities" (Cunsolo Willox et al., 2012). The pathways through which climate change impacts mental wellness in the Arctic varies by gender (Bunce et al., 2015, 2016; Harper et al., 2015) and age (Ostapchuk et al., 2015; Petrasek MacDonald et al., 2015, 2016). Emotional impacts of climate-related changes in the environment were significantly higher for women compared to men, linked to concern for family members (Harper et al., 2015), however, men are vulnerable due to gendered roles in subsistence and cultural activities (Bunce et al., 2015). In coastal areas, sea ice means freedom for travel, hunting, and fishing, so changes in sea ice affect the experience of place, which influences individual and collective mental/emotional, spiritual, social, and cultural health according to relationships between sea ice use, culture, knowledge, and autonomy (Gearheard et al., 2013; Canada, 2014; Durkalec et al., 2015; Inuit Circumpolar Council, 2015).

3.4.3.3.4 Infrastructure (transport, buildings, pipelines, life cycle costs)

Rural and urban

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

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

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

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

Ice roads

Winter roads (snow covered ground and frozen lakes) influence the reliability and costs of transportation to supply and connect remote northern communities and industrial development sites (Parlee and Furgal, 2012; Huskey et al., 2014; Overland et al., 2017a). For travel to and between northern communities, changing lake and river levels and the period of safe ice cover all affect the duration of use of overland travel routes and inland waterways, with associated implications for increased travel risks, time, and costs (Laidler, 2012; Ford et al., 2013; Goldhar et al., 2014). Reductions in ice cover duration and ice thickness (Section 3.4.1.2.1) create problems of accessibility for northern communities by reducing the reliability of traditional ice-based routes and safety of ice-based travel. As the ice becomes less predictable due to climate change, the connections between settlements becomes more difficult to maintain.

The reliability and predictability of ice roads as supply lines to northern development sites is not as dependent on climate driven sensitivity to ice conditions because these ice roads are managed each season. Ice growth is accelerated by removing overlying snow and flooding with lake water. Nevertheless, there have been recent instances of severely curtailed ice road shipping seasons due to unusually warm conditions in the early winter, which prevented sufficient ice growth to allow intervention (Sturm et al., 2017). While the impact of human effort on the seasonal development and maintenance of ice roads is difficult to quantify, it is projected that the Tibbitt to Contwoyto Winter Road, which spans 400 km primarily across frozen lakes within the Northwest Territories of Canada, will experience a reduction in the operational time window due to winter warming (Mullan et al., 2017).

3.5 Responding to Climate Change in Polar Systems

3.5.1 Introduction

Human responses to climate change in Polar Regions (like other regions) are part of a social process of actions taken concurrently at multiple levels – by individuals, households, communities, regions, nation states, and the international community. In this respect, human perceptions and choices, be they by a family, decision makers of all levels of government, a local community, national-level agencies, or the Arctic Council (AC), all constitute dimensions of human responses. In the complex milieu of Polar Regions, the social, cultural, economic, political and legal systems that shape responses of actors interact. Similarly, climate change interacts with other forces for change (e.g., resource extraction, land-use change, economic change), which necessitates a consideration of interacting forces, the potential for cumulative effects, and pathways forward that build resilience (Nymand and Fondahl, 2014; Arctic Resilience Report, 2016) *(high confidence).*

3.5.2 The Polar Context for Human Responses to Climate Change

Polar Regions represent one extreme of a continuum of the earth's social-ecological systems, with human responses to climate change occurring in unique and challenging conditions. In both northern and southern high latitudes, extreme climatic conditions and remoteness from densely populated regions constrain human actions, with constraints the result of restricted human mobility, limited opportunities for ecological productivity and renewal during warmer seasons, the paucity of baseline data, difficulties and high cost of logistics, and a complex geo-political milieu. A diversity of stakeholders with different economic interests and cultural orientations, including traditional indigenous hunters and fishers, non-indigenous and rural residents, international mining and oil and gas corporations, wilderness advocates, tourists, and polar scientists, also challenge collective responses to climate change, (Shadian, 2014; Shadian, 2017) *(high confidence).*

Humans are relatively recent arrivals in Antarctica, and for the most part are transient research scientists, and commercial fishers, tourism operators and one-time tourists. In short, there are no 'citizens' of the Antarctic and it is no one's homeland. The northern latitudes, on the other hand, have for millennia been the homelands of indigenous peoples, who today reside side by side with non-indigenous immigrants. Of the approximately 4 million people who reside in the Arctic, about 10% are counted as Indigenous, although determinations of what constitutes "indigenous" are disputed (Arctic Resilience Report, 2016). The composition of populations do vary by region. For example 85% of Nunavut, Canada are indigenous, in areas of Siberia only 2% are indigenous people, and about 15% of Alaskans are Native (Fondahl et al., 2015). Ethnicity and cultural orientation shape climate change responses (Adger et al., 2012) All arctic residents, however, are under the jurisdiction of southern based national-level states that hold sovereignty of Arctic lands and waters, although Greenland now has "home rule" with Denmark and settlement agreements with indigenous people in regions of North America give provide for levels of self government.

Consequently, human responses to climate change by arctic residents are significantly affected by histories of colonization and political relations in which southern-based nation states dictate and impose policies (Keil and Knecht, 2016). Human responses to climate change in Antarctic, on the other hand, are largely shaped by international agreements (*high confidence*).

In all regions of the Arctic, indigenous societies continue to be sustained with mixed cash-subsistence economies that are highly dependent hunting, herding, fish, herding, and gathering (Nymand and Fondahl, 2014). This high dependence on and relationship with ocean and land living resources, indigenous people's long history with their homelands, and the strong climate signal in the Arctic give indigenous peoples a sensitivity to climate change that informs their understanding of adaptation (Ford et al., 2015; Pearce et al., 2015) and approaches resource governance (Danielsen et al., 2014). In this respect human responses to climate change are considered by many as a matter of cultural survival (Greaves, 2016)(See Cross Chapter Box 3) *(high confidence)*. Indigenous people, however, are not apart from other sectoral activity areas. While in some cases they are impacted by sectoral activities negatively (Nymand and Fondahl, 2014), in other cases they benefit financially (Shadian, 2014), which at times creates dilemmas and conflicts (Huskey, 2018; Southcott and Natcher, 2018)*(high confidence).*

Both the Arctic and Antarctic are also unique with respect to the novelty of their systems of governance. The Antarctic Treaty, indigenous land claims and self-governance agreements, the role of Sami Council in Fennoscandia, Russian Association of Indigenous Peoples of the North in Russia, Inuit Circumpolar Council, nature-based NGOs, various resource co-management arrangements, and the Arctic Council are a few of the governance innovations in Polar Regions that make for a complex environment of decision making, with make multi-level linkages providing opportunities for humans to respond to climate change *(medium confidence).*

3.5.3 Characteristics of Highly Resilient Polar Systems

Recent development in literature on Polar climate change assessments has, to some extent, shifted from a sole focus on vulnerability to an examination of social-ecological resilience and pathways for building resilience, including the potential for adaptive governance. This shift represents an effort to capture the dynamic nature of social-ecological change, the potential for social-ecological system regime shifts, and the tremendous role of human agency in adaptation and transformation (ARIR, 2013; Arctic Resilience Report, 2016; AMAP, 2017a). Since adapting to the effects of climate change, communities, enterprises and institutions can build up their resilience to climate, this focus also begs the questions - How might climate induced regime shifts affect human well-being? Are some individuals and groups of Polar Regions more able to adapt and or transform to climate change than others? What conditions contribute to or impede adaption and transformation (AMAP, 2017a; AMAP, 2017a), and what choices are critical when and by whom to realize resilient climate pathways for the future?

Scholars of social-ecological systems have identified several sets of principles (i.e. conditions or system properties) that contribute to the general resilience of social-ecological systems (Chapin et al., 2010a; Chapin et al., 2010b; Biggs et al., 2015; Quinlan et al., 2016). They also provide a basis for assessing human responses to climate change. Biggs et al. (2012); Biggs et al. (2015) identified properties enhancing resilience of ecosystem services as i) high diversity and redundancy, ii) use of a complex systems approach to understand phenomena and address problems, iii) horizontal and vertical linkages between system elements, iv) social learning, v) broad participation in decision making, and vi) consideration of slow and fast variables when making decisions contribute to resilience. These principles can serve to evaluation both past actions and guide future responses. Table 3.3 lists these conditions with examples from the polar context.

Adaptive capacity to climate change in the Polar Regions, be it proactive, incremental, or transformative, is related directly to a group's access to tangible and intangible resources (Hovelsrud and Smit, 2010; Kofinas et al., 2013; Kofinas et al., 2016; Berman, 2017), as well as the group's motivation / empowerment to make use of those resources with action (Table 3.4). The importance of particular resources is context specific; while some resources may be especially critical for a group facing one change (e.g., reindeer herders responding to rain-on-snow events on the Yamal Penn of Siberia need movement to access to other pastures), those needed by another group will differ (e.g., commercial fishers of Tromsø responding to changes in the distribution and abundance of fish stocks may need additional financial resources for fuel and responsive and sympathetic systems of resource management.) Hence, appraising a group's access to the resources needed for adaptation is part of an assessment of adaptation *(medium confidence).*

Adaptive capacity has also been a topic of much recent academic review (Kofinas et al., 2013; Ford et al., 2014; Pearce et al., 2015; Berman, 2017; Gerlach et al., 2017). Ford et al. (2015) note that there is currently limited understanding of when and how Arctic adaptation takes place, and Berman (2017) argue that conceptual inconsistencies of terms used in different case studies of arctic adaptation limits the development of theory on adaptation and the applications of theory to policy. Others have noted the limited attention to environmental justice issues of adaptation, which do not take fully account for the hardships endured by some less empowered and endowed people when having to respond (McCauley et al., 2016; Huntington et al., 2018). How sectors respond, as presented below (Section 3.5.4), provides a basis for understanding the context-specific nature of adaptation among Polar actors. There remains a knowledge gap in the relationship of adaptation resources to context and action represents an area worthy of investigation (Kofinas et al., 2013; James et al., 2014; Berman, 2017) *(high confidence.)*

Table 3.3: System properties contributing to resilience in the Polar context

Conditions contributing to resilience (Biggs et al 2015)	Polar context application of characteristic
Diversity and redundancy	The high number of culture groups represented in the arctic, providing different perspectives on change; Multiple stakeholders who have shared "ownership" of a region
Use of a complex systems approach to understand phenomena and problems	Increased number of interdisciplinary efforts to analyse problems which provide a more holistic understanding of change and their implications and responses to change;
Horizontal and vertical linkages between system elements	Institutional arrangements that link local, regional, national, and international levels as part of governance that help with problem definition and coordination of responses
Social learning	National and international initiatives that inform shadow networks and sectoral decision making through intentional processes that facilitate reflexive action
Participation in decision making	Established expectation in many regions of the arctic that research and decisions include local voice, including consideration of local and traditional knowledge.
Consideration of slow and fast variables when making management decisions	A focus on underlying "slow" variables that govern long-term system behavior a cascading effects of change on various trophic levels, including implications to humans

Table 3.4: Assets (or resources) for adaptation with examples based on arctic subsistence harvesting

Asset for Adaptation	Example based on Arctic Subsistence harvesting
Geography	Proximity to needed and valuable resources
Ecosystems	Resource diversity that allows for species switching in needed; a system that rebounds after disturbance
Physical infrastructure	Roads for access, gear, internet for communication with outside entitites
Human capital	Skills to advocate for policy changes that address community needs
Social and cultural capital	Trust relationships internal to community and external with those who may work with community (e.g., researchers); social networks that provide links to information and other resources
Institutions	Formal co-management arrangements that give a community level of authority in policy making, and which responds quickly to concerns and changes
Financial capital	Access to funds to hire human resources, purchase new gear, buy gas for travel
Knowledge	Baseline data for retroactive studies, traditional and local knowledge on understanding of ecosystem practices associated with subsistence

3.5.4 Responses by Sector

Table 3.5 summarizes the consequence of climate change for various sectors as noted in Sections 3.3.4 and 3.4.3.3, their documented responses, key assets identified as supporting their response, and other drivers of change that could potential interact with climate change. The sector responses assessed here is not exhaustive; it does include important activity areas of Polar Regions relevant to climate change.

Table 3.5: Consequences, interacting drivers, responses, assets of climate change responses by sector.

Dimension/sector	**Consequence of climate change**	**Documented responses**	**Key assets and strategies identified as supporting adaptive and transformative capacity**	**Other drivers of change that may interact with climate and affect outcomes.**
Commercial Fisheries	Consequences are complex, affecting abundance and distribution of different fish 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	Changes in human preference, demand, and markets, changes in gear, changes in policies affecting property rights.
Subsistence (marine and terrestrial)	Changes in distribution and abundance with 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).	Change in gear, timing of hunting, species switching;	Systems of adaptive co-management that allow for species switching, changes in harvesting methods and timing, secure harvesting rights.	Changes in cost of fuel, land use affecting access, food preferences, harvesting rights
Reindeer Herding	Rain-on-snow events causing high mortality of herds; shrubification of tundra pasture lowering forage quality	Changes in movement patterns of herders; policies to insure free – range movements.	Flexibility in movement to respond to changes in pastures, secure land use rights. Continued economic viability and cultural tradition.	Change in market value of meat; overgrazing; Land-use policies affecting access to pasture and migration routes, property rights
Non-Renewable Resource Extraction	Reduced sea ice and glaciers offering some new opportunities for development; changes in hydrology (spring run off), thawing permafrost, and temperature affect production levels, ice roads, flooding events, and infrastructure	Some shifts in practices, greater interest in off shore and on-land development opportunities in coms regions.	Modification of practices and use of climate change scenario analysis.	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)

Transportation	Open seas allowing for more vessels; greater constraints in use of ice roads	Increase shipping, tourism, more private vessels	Strong national and international cooperation leading to agreed upon and enforced policies that maintain standards for safety; well development response plans with readiness by agents	Political conflict in other areas that impeded acceptance of policies for safety requirements, timing, and movements.
Infrastructure -urban and rural human settlements, year-round and winter roads	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.	Weak regional and national economies, other disasters that divert resources, disinterest by southern-based law makers
Coastal settlements (See Cross Chapter Box 5: Low-lying Islands and Coasts)	Change in extent of sea ice with more storm surges and thawing of permafrost, to 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.	Limitations of government budgets, other disasters that may take priority for spending, deficiencies in policies for addressing mitigation and relocation
Tourism (Arctic and Antarctic)	Warmer conditions, more open water, Public perception of "last chance" opportunities,	Increased visitation, increase in off-season tourism to polar regions	Policies to insure safety, cultural integrity, ecological health	Travel costs. Shifting tourism market, more enterprises
Human Health	Threats to food security, potential threats to physical and psychological well being	Greater focus on food security research; programs that address fundamental health issues	Human and financial resources to support public programs in hinterland regions; cultural awareness of health issues as related to climate change.	A reduction (of increase) in public resources to support health services to rural community populations, research that links ecological change to human health

3.5.4.1 Fisheries

In the Arctic globally significant and economically valuable commercial fisheries take place in the Barents, northern Norwegian and Bering Seas (Alaska Fisheries Science Center, 2016). Seasonal and interannual variability in ocean conditions influences product quality, quantity and catchability (Haynie and Pfeiffer, 2012). As documented in Section 3.3.3, climate change is expected to impact the spatial distribution and productivity of marine fish and shellfish in different ways depending on the vulnerability (a combination of exposure and sensitivity) of the species and specific stock. As also documented in Section 3.3.3 climate change is impacting the balance between subarctic and high Arctic communities, with unclear effects on future fisheries.

The North Pacific Fishery Management Council currently employs an ecosystem approach to fisheries management in the southeastern Bering Sea groundfish fishery (Livingston et al., 2011) and efforts to explore the performance of the current system under changing climate conditions is employing scenario-informed management strategy evaluations to inform management (Holsman et al., 2017). The existence of science based holistic management strategies in the southeastern Bering Sea portends that the management of marine resources in the Arctic will be founded in precautionary approaches to sustaining marine resources and ecosystem structure to the extent possible. The fisheries of the southeastern Bering Sea are managed sustainably through a complex suite of regulations that include catch shares (Ono et al., 2017), habitat protections, restrictions on forage fish, bycatch constraints (DiCosimo et al., 2015), and community development quotas. These complex interacting management constraints are consistent with an ecosystem approach to fisheries management (Dolan et al., 2015). This intricate regulatory framework has inherent risks and benefits to fishers and industry by limiting flexibility (Anderson et al., 2017b).

In 2009, a new Marine Resources Act entered into force in Norway, including Norwegian sectors of the Arctic (Barents and Norwegian Sea). The new act applies to all wild living marine resources and states that its purpose is to ensure sustainable and economically profitable management of the resources. Conservation of biodiversity is described as an integral part of sustainable fisheries management and it is mandatory to apply "an ecosystem approach, taking into account habitats and biodiversity" (Gullestad et al., 2017). Beyond national management the Joint Norwegian-Russian Fisheries Commission provides joint 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 of the annual negotiations between Norway and Russia. Since the turn of the century, the Commission has been working towards a long-term, precautionary approach to harvesting strategies. For example, the code of conduct for quota stipulation takes into account that capelin, with a short life span and fluctuations in stock biomass according to environmental conditions, is the main food source for cod. The total capelin quota is established such that it allows for a 95% probability that at least 200,000 tons of capelin spawn every year (http://www.jointfish.com/eng.html).

If managed properly Arctic fisheries need not be negatively impacted by moderate future warming (European Parliament's Committee on Fisheries, 2015). For example, the Norwegian cod fishery has exported dried cod over an unbroken period of more than thousand years (Barrett et al., 2011), reflecting the resilience of the northern Norwegian cod fisheries to historic climate variability (Eide, 2017). The high present yield of the Barents Sea (Section 3.3.3.1) and model projections indicate that enlarged habitat and increased production of plankton and prey due to increasing temperatures and ice retreat, may ensure that the migratory fish stocks remain large and the economic benefits from fisheries continues (Lam et al., 2016; Eide, 2017).

Five nations have existing EEZs in the high Arctic and each nation manages their resources within the regulatory measures of their nation. A review of future harvest of living resources in the European Arctic by Haug et al. (2017) points towards high probability of increased northern movement of several commercial fish species (Section 3.3.3.1 and Box 3.3), but only to the shelf slope for the demersal species. This suggests increased northern fishing activity, but within the 200 nm zones and the present management system (Haug et al., 2017).

Commercial fishing is currently prohibited in the US portions of the Chukchi and Beaufort Seas (Wilson and Ormseth, 2009). In the Canadian sector of the Beaufort Sea commercial fisheries is until now only small scale and locally operated, but climate change with decreasing ice cover together with over-harvesting of fish stocks other places may increase the incitement. This has caused concern among local Inuvialuit subsistence fishers and a new proactive ecosystem-based Fisheries Management Framework was developed (Ayles et al., 2016). In 2015, the Oslo declaration on high seas fishing in the central Arctic Ocean was signed which established a moratorium on commercial fishing in the central Arctic Ocean and encouraged research cooperation amongst the bordering nations. These constraints will limit the expansion of commercial fishing until sufficient information if available to sustainably manage fisheries under the influence of climate change.

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

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

3.5.4.2 Subsistence Economies of the Arctic (Marine and Terrestrial)

Subsistence in the Arctic is non-market hunting, fishing, and gathering, and involves a set of social activities - preparation for the harvest, traveling on the land or sea, harvest, food preparation, and sharing of the take, and giving thanks to harvested animals. Subsistence in the Arctic is part of a mix-cash economy, with the balance differing by region. Rural community residents of the arctic represent the greatest proportion of subsistence users and are the most dependent on wildfoods, but urban residents are also involved to varying degrees. Subsistence occurs at many levels (individual, household, community, regional), with the household being a central unit. Thus, human responses to climate change by those who are part of subsistence encompasses many aspects of life.

The recent AMAP Adaptation Actions for a Changing Arctic (AACA): Perspectives from the Bering-Chukchi-Beaufort Region (AMAP, 2017a), and Perspectives from the Barents Region (AMAP, 2017b) provide many rich details on human responses to climate change. As reported in the AACA, responses fall into several categories. In the area of harvest preparation, harvesters are adjusting their assessments of safety in response to changing conditions and an increase in the number of accidents due to unsafe ice. They are also shifting the timing of harvesting and the selection of harvest areas due to changes in seasonality. Consequently there is a greater number of no-go days, which can affect overall harvest success. Changes in the navigability of rivers and more open (i.e. dangerous) seas has resulted in harvesters changing harvesting gear, such as shifting to from propeller to jet-propelled boats or all-terrain-vehicles, and to larger ocean-going vessels for traditional whaling. In both cases, the newer gear results in an increase in fuel costs (e.g., jet boats are about 30% less efficient) (Kofinas et al., 2010; Brinkman et al., 2014). Because of the many factors affecting the abundance and distribution of species (e.g. caribou), attributing changes in harvest success is difficult. Clearly, caribou hunters where herds have dramatically decreased (Section 3.4.3) have responded with species switching or doing without. Evidence and personal accounts, however, do show indicate that in many cases, harvesters' adaptive responses have allowed for continued success in the provisioning of wildfoods (BurnSilver et al., 2016; Fauchald et al., 2017; AMAP, 2017a)(*medium confidence*).

People are responding in other activity areas subsistence as well. Difficulties drying wildfoods (e.g., fish) because of increased summer precipitation and more cloudy days, and the thawing of ice cellars has led to an increased use of household and community freezers, and in some cases an abandonment of traditional food drying practices. New technology to mitigate ice cellar thawing is being tested in several villages of the North Slope of Alaska. And some cases there has been an increase emphasis on community self-reliance such as use of household and community gardens for food production (Loring et al., 2016). It has also spawned new programs to build partnerships with university researchers who are interested in studies that meet community needs (Chapin et al., 2016), and an overall greater willingness to engage in knowledge co-production activities. Because Arctic residents are aware of the sources and impacts of climate change at a global scale, there are also efforts to use alternative energies in remote villages, such as solar panels. To be sure, the motivation to pursue these technologies is to a great extent economic, but they are also motivated by a perceived global crisis (*medium confidence*).

The perception of a global climate crisis and the need for cultural survival have resulted in indigenous people engaging in political processes on climate change at many levels and in different avenues. At the United Nations Framework Convention on Climate Change (UNFCCC), the discursive space for incorporating various 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). Through the AC, Aleut International Association, Arctic Athabaskan Council, Gwich'in Council International, Inuit Circumpolar Council, and Russian Association of Indigenous Peoples of the North, and the Saami Council sit as permanent participants, and are involved in many of the AC's working groups. (Sections 3.5.5.2.2 and 3.5.5.2.3). Greater involvement at the national and regional levels has also occurred through the structures and provisions of indigenous settlement agreements (e.g., Nunavut Act, 1993), fish and wildlife co-management agreements, and participation in boundary organizations. For example, changes in seasonality have led hunters to propose changes in wildlife management regulations on the timing of moose hunting seasons (*high confidence*).

The boundaries between impacts, responses, and outcomes of social costs and social learning are linked, with involvement in such processes coming at the expense of high transaction costs (i.e., greater demands on overburdened indigenous leaders, and adding to stress of communities living with limited resources. The threat of climate change can also reinforce cultural identify and experiences of political involvement can allow indigenous leaders to be more effective agents of change in these arenas (*low confidence*). Penn et al 2016 point to these issues, arguing the need for a greater focus in assessments on community capacity and cumulative effects.

3.5.4.3 Reindeer Herding

Herders' responses to climate change vary by region and their respective herding practices (Klokov, 2012; Forbes et al., 2016; Uboni et al., 2016; Mallory and Boyce, 2017). For example in Fennoscandia husbandry practices of reindeer by some (mostly Sami) include supplemental feeding, which provide buffer for unfavorable conditions. In Alaska, reindeer herding is primarily free range, where herders respond to icing events and potential loss of reindeer because movements of caribou herds (wild reindeer), both of which are partially driven by climate.

With Nenets of the Yamal, their resilience in herding has been facilitated through herders' own agency and, to some extent, the willingness of the gas industry on the Yamal to observe non-binding guidelines that provide for herders' continued use of traditional migrations routes (Forbes et al., 2015). In response to climate change (i.e., icing events and early spring run offs blocking migration), the only way of avoiding high deer mortality is to change the migration routes or take the deer to other pastures. In practice, however, the full set of challenges has meant more nomadic herders opting out of the traditional collective migration partially or entirely to manage their herds privately, or ending herding as a livelihood. 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 pasture conditions as infrastructure expands (Forbes, 2013) The same logic has more recently been applied in the wake of recent rain-on-snow events (Forbes et al., 2016) *(high confidence).*

3.5.4.4 Non-renewable Extractive Industries

Activities of non-renewable resource extraction are determined by several factors, such as treaties and national-level polities, global hydrocarbon and mineral markets, and cost of operations, with the later potentially affected by a changing climate and the former contributing to climate change. Exploitation of natural resources in the Antarctic is prohibited by the Antarctic Treaty. In the Arctic, receding sea ice and glaciers has opened new possibilities for development, such as areas of receding glaciers of eastern Greenland (Smits et al., 2017). As exploration got underway in Greenland, its home rule Government began developing environmental impact assessment protocols to provide for adequate public participation (Forbes et al., 2015). On the North Slope of Alaska, oil and gas development is now undergoing new growth, while industry concurrently faces increasing environmental challenges, some of which are related to shorter and warmer winters, the main season for oil exploration and production (Lilly, 2017). Lilly reports that optimizing North Slope transportation networks in oil field during winter operation seasons is critical in managing increasing resource development, and could provide a framework for environmentally-responsive development. Better understanding of environmental change is also considered important in insure continued oil field operations with protection natural resources. Better forecasting of short-term conditions (snow, soil temps, spring run offs) could allow management agencies to respond to conditions, and industry more time to plan winter mobilization. looding events on the North Slope of Alaska due to unusually high spring melt and run off in 2015 closed the Dalton Highway and North Slope oil field operations for an extended period, resulting in financial losses to companies, and suggesting that industry rethink the design of culverts and roads (Raynolds et al., 2012). This assessments is limited because of there is few peer-reviewed literature on responses by industry to climate change *(low confidence).*

3.5.4.5 Infrastructure

Humans are now facing issue of road flooding, coastal and river erosion at settlement locations, and damage to building foundations, runways, pipelines, and other forms of infrastructure, occurring in many Polar Regions as a consequence thawing permafrost, sea ice retreat, and changes in hydrology (AMAP, 2017a; AMAP, 2017b). Melvin et al. (2017a) estimated costs damages (without adaptation measures) from 2015 to 2099 will be $4.2billion to $5.5 billion, depending on the climate scenario. Estimates of proactive adaption measures (i.e., reduction of greenhouse gases) is estimated to reduce damage costs by half. Impacts in urban infrastructure in Arctic Russia are likely to be high, given its high number of large settlements and the dependence of the Russian economy on sectors based in northern regions, requiring a triage assessment when responding. Regional- to local-level adaptation measures are affected by budgetary constraints, while maintenance of infrastructure (e.g., road maintenance) has already increased operating costs for local and regional governments. In Alaska were the state is facing economic hardships, state funding previously allocated to support coastal community relocation was reallocated to address flooding events in another part of the state (x). Method for building of ice roads on the North Slope have been somewhat modified to account for warmer temperatures during construction *(low confidence).*

3.5.4.6 Tourism

Climate and weather are critical considerations in tourism planning and development (Saarinen, 2014). There has been a growth in tourism in both Polar regions, with some firms capitalizing on ‘last chance tourism’ perceptions (i.e., “see it before its gone”) (Lamers et al., 2013). The growth of this sector market is anticipated to increase in near- and later-term future especially with the travel of small vessels (yachts) (Johnston et al., 2017). Polar-class expedition cruise vessels are now, for the first time, being purposefully built for recreational arctic sea travel. Opportunities for tourism vessels to contribute to international research activities (‘ships of opportunity’), may improve sovereignty claims in some regions, contribute to science, and enhance education among public about Arctic regions (Stewart et al., 2013; Arctic Council, 2015; Stewart et al., 2015). The anticipated grow of cruise tourism in Polar Regions also points to the need for operators, governments, destination communities, and others to identify and evaluate adaption strategies, such as disaster relief management plans, updated navigation technologies for vessels, codes of conduct for visitors, and improved maps (Dawson et al., 2016). As well, limited research has examined perceptions of tourism and appropriate adaptation responses by residents of local community destinations (Kaján, 2014; Stokke and Haukeland, 2017). Efforts were initiated with stakeholders in Arctic Canada to identify strategies that would lower risks, but additional research is needed to evaluate empirically strategies (Dawson et al., 2016)*(medium confidence).*

Tourism activities in the Antarctic have also increased considerably. The industry organization that manages much of the tourism activity in Antarctica, the International Association of Antarctic Tour Operators, has been working with Antarctic Treaty Consultative Parties to manage changes in operations and their impact on ice-free areas (ATCM, 2016). It has been suggested that use of existing protected area management mechanisms should be used to mitigate some of the impacts of high visitation rates (ASOC, 2015). However there is a general disagreement about the regulation of Antarctic tourism among Treaty Parties and the benefits parties derive from tourism are currently not shared. Climate change is a challenge because it is often considered as an external factor that can be dealt with from a scientific perspective. The focus on climate change science has drawn attention away from the development of normative responses to assessing and managing human activities in the Antarctic in a responsible way, such that the human footprint and greenhouse gas emissions are minimized. Legal basis applying are the Madrid protocol (Art. 3) requiring a minimization of adverse environmental impacts vs. global environmental regimes (such as ATS) to a greater extent. (Dodds, 2010; Hemmings and Kriwoken, 2010; Orheim et al., 2011; Triggs, 2011)*(medium confidence).*

3.5.4.7 Transportation

Arctic shipping activity, especially in certain geographic areas (NSR, AB and eventually NWP and maybe TPR) has and is likely to continue increasing in the future (Stephenson et al., 2011; Smith and Stephenson, 2013; Stephenson et al., 2013). These increases are occurring in spite of the limited total savings when comparing shorter travel to increased CO_2 emissions (Lindstad et al., 2016). Without developed and refined risk management plans, increases in traffic will result in greater risk to humans and ecosystems, such as the introduction of invasive species (Ware et al., 2014) and oil spills. For example Statoil has developed and uses of risk assessment decision-support tools for environmental management, together with environmental monitoring (Utvik and Jahre-Nilsen, 2016). The tools allow for qualification to assess Arctic oil-spill response capability, ice detection in low visibility and improved management of sea ice and icebergs, and numerical modelling of icing and snow as risk mitigation. In anticipation of spills, research in several regions have explored oil spill response viability and new methods of oil spill response for the Arctic environment (e.g., Bullock et al., 2017; Dilliplaine, 2017; Holst-Andersen et al., 2017) *(medium confidence).*

The International Maritime Organization is the organization responsible for international Arctic shipping regulating. There are a number of mechanisms standardizing regulation and governance (MARPOL, SOLAS, STCW.), including recent Arctic initiatives, such as joint search and rescue agreements and joint oil pollution response, and the newly implemented Polar Code (IMO, 2017). The Polar Code does address emerging issues, but is likely to need additions and modifications in the future; the agreement was consensus based, hence implemented at the lowest common denominator, including a call to enhance enforcement capabilities and address emerging issues such as heavy fuel oil and black carbon, among other environmental protection provisions regulating heavy fuel oil (HFO) transport and use, black carbon, and ballast water (Anderson, 2012; Sakhuja, 2014; IMO, 2017). National-level regulation varies (some stronger than others) and ships with flags of convenience can cause challenges (Chircop, 2009; Anderson, 2012; IMO, 2014; Sakhuja, 2014). Continued international cooperation on shipping governance is needed (Arctic Council, 2015; Arctic Resilience Report, 2016; PEW Charitable Trust, 2016; Chénier et al., 2017; IMO, 2017). National-level responses have included several studies to consider scenarios of change and explore regulatory changes *(high confidence).*

Many airstrips in Antarctica are built on ice. As demonstrated by the ice airstrip near the Italian Mario Zuchelli station, the safety and ability of landing of aircraft on the airstrip has been compromised by the changing climate (Italy, 2015). Due to this impact on its transportation access to its station, Italy has proposed to construct a gravel runway (Italy, 2015). With frequent use of ice runways and increase of air traffic by both National Antarctic Programmes and tourism operators, unpredictable states of airstrips will alter such transportation and the infrastructure to support them (ATCM, 2017). The the IMO Polar Code came into force in 2017 with the purpose of setting new standards for vessels. travelling in polar areas to avoid environmental damage and to improve safety (IMO, 2017). The IMO Polar Code, however, currently excludes fishing vessels and vessels on government service, thereby excluding many shipping activities in the Antarctic region (IMO, 2017)*(high confidence).*

3.5.4.8 Arctic Human Health and Well being

The stress that climate change places on health systems is particularly concerning for Polar Regions (Watts et al., 2015), where health systems already face the challenges of remote service provisioning, including weather-dependent modes of transportation for supplies and services; high costs of delivery; high existing burdens of disease; access to health services in remote locations; high turnover of health professionals; and an identified need for increased culturally-appropriate health programming (Chatwood et al., 2010; Minore et al., 2004; Wexler et al., 2008; Young et al., 2011). The ability to manage, respond, and adapt to climate-related health challenges in the future will be a defining issue for health sector in the Polar Regions (Blashki et al., 2011; Cunsolo et al., n.d.; Sibbald, 2013) *(high confidence).*

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

Many health adaptation efforts by governments have been groundwork actions, focused increasing awareness of the health impacts of climate change and conducting vulnerability assessments (Austin et al., 2015; Lesnikowski et al., 2011; Panic et al., 2013). For instance, in Canada, this has included training, information resources, frameworks, general outreach and education, and dissemination of information to decision makers (Austin et al., 2015). Finland's federal adaptation strategy outlines various anticipatory and reactive measures for numerous sectors, including health (Gagnon-lebrun et al., 2007). However, an increasing number of government adaptation actions have also taken place, which are aimed to reduce vulnerability, including warning and monitoring systems, as well as initiatives aimed at changing practice and behaviour. For instance, all Polar Regions in Northern Canada have developed climate change adaptation plans, within which health sector initiatives are outlined (Austin et al., 2015), and there is federal health programing designed to build Arctic health-related climate change adaptive capacity and promote adaptation action (Peace et al., 2012). In Alaska, the Arctic Investigations Program responds to infectious disease via advancing molecular diagnostics, integrating data from electronic health records and environmental observing networks, as well as improving access to in-home water and sanitation services. Furthermore, circumpolar efforts are also underway, including an circumpolar working group with experts from public health to assess climate-sensitive infectious diseases, and to identify initiatives that reduce the risks of disease (Parkinson et al., 2014). Importantly, health adaptation is occurring at the local scale in Polar Regions (Ford et al., 2014a; Ford et al., 2014b). The types of groundwork and action adaptation at the local scale is broad, from community freezers to increase food security, to community-based monitoring programs to detect and respond to climate-health events, to Elders mentoring youth in cultural activities to promote mental health when people are "stuck" in the communities due to unsafe travel conditions (Austin et al., 2015; Bunce et al., 2016; Cunsolo Willox et al., 2017; Douglas et al., 2014; Harper et al., 2012; Pearce et al., 2010) *(high confidence).*

3.5.5 Cooperation in Multilevel Governance of Polar Climate Change

The nexus of discourse on international cooperation and responses to climate change in Arctic environmental governance is changing. Rather than treating Polar transformations and their governance in isolation (i.e. purely with regional lens (Cassotta et al., 2016; Keil and Knecht, 2017)), polar systems require assessments in a global multi-regulatory levels fashion that links global climatic, environmental and political systems and environmental processes in a triadic nexus between impacts, human adaptation and indigenous knowledge. Regions are today responding to climate change through governance across different levels and sectors. Hence, Polar responses are assessed within the context of synergistic linkages and reverberations between different pluralistic levels of governance, and with different institutional arrangements, both formal and informal, involving a set of formal and informal actors and networks operating with different norms.
The new landscape of Polar governance is a fragmented spectrum of sources of law and policy at global, regional, national and local levels, that moves from a vertical and horizontal interplay (Molenaar, 2012; Koivurova, 2016; Young, 2016)(*high confidence*). Vertical coordination is required increasingly between the different levels of governance, and horizontal coordination between regulatory sectors, so that the problems can be tackled more effectively by exploiting such synergies. Vertical coordination is necessary for the implementation of law and policies in terms of environmental effectiveness, management and public participation, especially for Arctic indigenous people involved governance processes. Our assessment is conducted though the lens of legal pluralism and polycentrism applied to Polar environmental governance, since multi-level governance builds a robust framework to assess interconnections and comparisons on the legal and political space (Stokke, 2009; Tuori, 2011; Young, 2011; Cassotta, 2012; Prior, 2013; Shibata, 2015; Cassotta et al., 2016) (*high level of confidence*, *robust agreement*). Primarily we examine ocean

governance with references to the cryosphere as is relevant. Within this complex multi-level, governance landscape, we focus on those Polar conventions and institutions that have a dual role in providing for cooperation as well as the triadic nexus between impacts, human adaptation and indigenous knowledge.

3.5.5.1 Formal Arrangements: Polar Conventions and Institutions

Both in the Arctic and the Antarctic, cooperation is strong, and takes place in different levels and intersecting sectors of multi-regulatory governance. At the global level one finds several instruments of cooperation, as well as at the international, regional level, at sub-regional level and at the national level (Koivurova, 2016; Young, 2016). All levels of sources of law and policy are interdependent. It is therefore important to consider the synergies and interactions across sectors and at different levels of governance where various actors are operating. An emblematic example of vertical implementation is given by the role of the United Nations Convention on the Law of the Sea (UNCLOS), a global convention which codifies customary international law. This convention implement agreements on biodiversity beyond national jurisdictions, which means that global conventions implement at national level *via* regional level, their own provisions. This is the case of UNCLOS implementing Annex V of the Convention on the Protection of Marine environment of North East Atlantic (OSPAR) a regional instrument (Jakoben, 2014), which is extremely relevant for example in the case of Arctic, for the protection against mining and oil and gas activities interacting negatively with climate change. The same Annex V implements the Convention on Biological Diversity (CBD) another crucial global convention that OSPAR is implementing at a regional level. However, OSPAR only applies to the North East Atlantic. Interactive interdependencies and synergistic linkages between the global-to-regional levels and vice-versa reverberate down at national level, a dynamism, which now characterises the new Polar Regions' regulatory landscape (Jakoben, 2014).This suggests the importance of synergies at different levels of cooperation, and most importantly, integration between the two different poles (Jabor, 2016).

3.5.5.2 Sectors of Cooperation and Responses in the Arctic

Cooperation in the Arctic mainly focuses on climate change, the (marine) environment, and scientific research. Cooperation is especially strong on climate change, especially through the role of the Arctic Council (AC). Currently, the issue of international cooperation in the sector of climatic issues focuses primarily on global problems, such as depletion of stratospheric ozone, the impact of persistent organic pollution and the effect of heavy metals. Recently, cooperation in dealing with climate change is relevant regarding the role of the European Union (EU), especially in safeguarding the Arctic environment. This is especially the case for European Arctic and Arctic-related economic development identified as "key priority areas" for the EU's action in the Arctic in the latest 2016 Joint Communication action (European Commission, 2016). The EU is an important source of financing for Arctic climate research, including Arctic reach projects, like the Horizon 2020 program running from 2014 (https://ec.europa.eu/programs). In the sector of marine environment, the future of Arctic governance, including the role of the AC, will also depend on the implications of the new recent agreement of on the Conservation and Sustainable use of Marine Biodiversity of Areas beyond National Jurisdictions (BBNJ), signed in December 2017 under UNCLOS. Cooperation is occurring also with scientific research stations via bilateral agreements such as the one between Norway and various countries dealing with scientific research operating at Ny-Ålesund in the Svalbard Archipelago (Goodsite et al., 2016).

3.5.5.2.1 Arctic Council

The AC is an example of cooperation par excellence and that is soft law nature, a new middle-way and *sui generis* meta-juridical institutional body operating in a new context of the climate change Arctic of globalization and transnationalism (Baker and Yeager, 2015; Cassotta et al., 2015; Pincus and Speth, 2015). In 2013 the AC granted China, South Korea, Japan, India, Italy and Singapore the status of permanent observer at the AC. Despite lacking the prominent role to enact hard law, the AC undertook the signature of the three agreements: 1) Agreement on Cooperation in Aeronautical and Marine Searched and Rescue in the Arctic, 2) the Agreement on Cooperation on Marine Oil Pollution, and 3) the very recent Agreement on Enhancing International Arctic Scientific Cooperation, are all signs that the AC is choosing to prepare to regulate climate changes in the Arctic using hard-law instruments. The role of the AC is gradually evolving especially in its work to increase knowledge about the circumpolar Arctic in order to influence both national and international policy-making. (Koivurova, 2016). A significant climate change related-report performed by the AMAP in 2017 is the "Snow, Water, Ice and Permafrost in the Arctic ("SWIPA 2017") portraying the status of the Arctic Cryosphere. It revealed accelerated change in major components of the cryosphere adopted by the 2011 Nuuk Declaration. This declaration put emphasis on the Arctic Cooperation with the goal of increasing Arctic adaptation and resilience, and enhancing the role of the AC as a leader to minimize the human and environmental impact of climate change (Tesar et al., 2016). The same Declaration established an expert group on Arctic ecosystem-based management and asked to review the need for an integrated assessment of multiple drivers of Arctic change and starts in what is defined "Arctic Change Assessment and Arctic Resilience Report (ARR) the final report of which has been delivered in 2016. In 2013 with the Kiruna Declaration, directives to undertake mitigation measures were established that led to the adoption in 2015 of the Framework for Action on Enhanced Black Carbon and Methane and Methane Emissions reductions under the 2015 Iqaluit Ministerial Declaration (Tesar et al., 2016). The latter declaration has focused on adaptation and resilience. The AC therefore has a) the potential role to enhance internal coherence in the current fragmented landscape of multi-regulatory governance in a complex system and b) potential chances to enhance its leadership role as integrator which would expect a reconstruction of its powers, c) potential role to regulate climate change Arctic trough hard law. The AC, however, is currently too small to deal with environmental global problems and operate in the transnational context.

3.5.5.2.2 Governance involving indigenous people: The ICC

The Inuit Circumpolar Council (ICC) is an international non-governmental organization representing about 160,000 Inuit living across the Circumarctic with a special consultative status with the United Nations Economic and Social Council (UNESC).The ICC was ECOSOC-accredited and was granted special consultative status (category II) at the UN in 1983, and has demonstrated its capacity to promote local Inuit governance into international and regional politics, as well as promote local-level adaptation to climate change. In terms of adaptation policies at global level, the ICC has urged global leaders to support its efforts by taking actions that commit them to include the participation of Inuit in climate-related processes and programs in a multi-level governance context (International, regional, national and local) by providing full effective participation in National Adaptation, Programs of Actions (NAPAs), the Disaster Risk Reduction (DDR) and National Adaptation Plans (NAPs), the Local Adaptations Plans for Actions (LAPAs) and the National Designation Authorities (NDAs) (UNFCCC). The ICC committed also to ensure that Inuit and Indigenous People have a voice in the Executive Committee of the Warsaw International Mechanism for Loss and Damage and full participation in and through the Adaptation Fund (UNFCCC). The ICC ensured that the Global Stocktake under the Paris Agreement included reports from community-based monitoring and information systems (CBMIS), as well as data and observations produced by principles and obligations of the Paris Agreement including a specific focus on human rights and the rights of indigenous people (ICC-UNFCCC). In terms of supporting adaptation at the regional level, the ICC works in conjunction with the AC's working groups to highlight the relevance and the need of investment in new infrastructure to assist Inuit communities to adapt to changes in climate, sea ice and shorelines. In their implementation of mechanisms of adaptation, states and the international community must commit to paying the costs of climate change adaptation measures, especially in upgrading the fuel-related infrastructure across Inuit Nunaat (Inuit homelands), and more broadly to assist northern communities to move away from carbon fuels. (Inuit Circumpolar Council, 2014).

3.5.5.2.3 New Trends and regulatory tools in adaptation governance with indigenous people

The vulnerability and resilience of climate change depends, to some extent, on institutions (i.e., rules, laws, policies) governing the use of resources. Community-based ecosystem monitoring, for example is potentially important in multi-level governance and regulatory processes; this approach recognizes locally-situated engagement with the environment and that governance should be informed by indigenous and local knowledge priorities. In collaboration with scientific research projects, for example, or monitoring for wildlife management, indigenous communities can be placed within wider international, regional and national networks, allowing local perspective on social-ecological change to be heard and local consent to be expressed through public participation. The Circumarctic *Rangifer* Monitoring and Assessment Network (CARMA) of CAFF is one such example (https://carma.caff.is/), which serves as a boundary organization. Assessing the many regulatory, strategic tools and potential strategies that could be used by non-state actors responding to climate change, especially in terms of adaptation including traditional knowledge, it should be noted that new forms of management of natural resources to comply with adaptation policies could be used by state or non-state actors, such as: co-management; indigenous protected areas; and indigenous corporate enterprises. Some examples of these new forms of strategic management and legal tools are the Best

Practices in Ecosystem-Based Ocean Management in the Arctic such as for example, those of Bepomar, Nunavut Wild Life Management Board (NWMB), Alaska Eskimo Whaling Commission (AEWC), Inuit Shipping, Inuit Airlines Companies and other Transportation Companies (Shadian, 2014). In addition, there will be new possibilities for indigenous people to work in partnership with resource developers (governments and local communities), which will open the path for using new business practices, and legal tools or strategies such as the use of Corporate Social Responsibility (CSR) or Impact Benefit Agreements (IBA) with indigenous people or a more extensive use of Environmental Impact Assessments (EIAs), Social Impact Assessments (SIAs) or Impact Benefit Agreements (IBAs) as a form of adaptation from a legal and business perspective (Forbes et al., 2015; Cassotta et al., 2016). The implementation of new practices in adaptive governance and practice at local levels reinforce the link between regional level (AC and ICC) and global level and may contribute in both reinforcing and establishing a new role of "indigenous internationalism" as a new mechanism of climate governance.

3.5.5.3 Sectors of Cooperation and Responses in the Antarctic

In the Antarctic, the variety of economically viable resources is limited. At present, the focus is on the only two economically viable resources: marine living resources and tourism. Currently cooperation does occur via UNCLOS, the Convention for the Safety of Life at Sea (SOLAS) and the Convention for the Prevention of Pollution from Ships (MARPOL) and the Polar Code, which applies to tourism vessels and through the IAATO managing of tourism in accordance to the ATS. Cooperation in the Antarctic also occurs with the CCAMLR falling under the ATS. Climate change is a big issue for the CCAMLR because it poses challenges regarding its impact on waters and the way to regulate and manage fisheries. Increasing water temperatures and higher acidification as a consequence of climate change aggravate the already big challenges in the Antarctic which are illegal, unreported and unregulated (IUU) fishing and the ensuing conditions in the marine environment (Jabour, 2017).

3.5.5.3.1 The Antarctic Treaty System (ATS)

The importance of understanding the impacts of changes to the Southern Ocean and Antarctic cryosphere, mitigating these impacts and adapting to them 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 (ATCPs) agreed that a Climate Change Response Work Programme would address these matters (ATCM, 2016), which led to the establishment of the Subsidiary Group of the Committee for Environmental Protection on Climate Change Response (SGCCR) (ATCM, 2017). The Commission on the Conservation of Antarctic Marine Living Resources (CCAMLR) has recognized the importance of climate change in its area of interest. As its last meeting (CCAMLR-XXXVI), however, CCAMLR was unable to agree a Climate Change Response Work Program (CCAMLR, 2017a).

3.5.5.4 Informal Arrangements

3.5.5.4.1 Networks and non-state actors in the Arctic

In the Arctic, indigenous people are considered as non-state actors. In many cases, but not always, they promote environmental protection in support of the sustainability of their traditional livelihoods, in opposition the pro-development business sector with its well-funded, strong power of lobbies. Indigenous people often face difficulties in these efforts because they only can negotiate with non-binding instruments of soft law, such as 1) The Circumpolar Inuit Declaration of Arctic Sovereignty of 2009 and 2) The Circumpolar Inuit Declaration of Resource Development Principles in Inuit of 2011. Bilateral agreements in the Arctic are typically state-owned (i.e. largely state dominated and controlled) enterprises, and are negotiated with powerful non-state actors, such as China National Petroleum Company; state-dominated companies such as Gazprom or Statoil and private corporations like Exxon Mobil (Young, 2016). Among the non-state actors, new economic forums have been established (Wehrmann, 2016). One example is the Arctic Economic Council (AEC), created by the AC during 2013-15 as an independent organization that facilitates Arctic business-to-business activities and is responsible for economic development (https://arcticeconomiccouncil.com).

3.5.5.4.2 Antarctic

The ATCPs, through the SGCCR, continue to work closely with the Scientific Committee on Antarctic Research (SCAR), the Council of Managers of National Antarctic Programs, the IAATO and other NGOs to understand, mitigate and adapt to impacts associated with changes to the Southern Ocean and Antarctic cryosphere. Various bilateral (e.g. IAATO-SCAR) and multi-lateral projects are underway to understand and to mitigate risk, many of these funded by national programs. Understanding, mitigating and adapting to climate change are among the key priorities identified for research in the region (Kennicutt et al., 2014a; Kennicutt et al., 2014b), and much has been done to understand how best to support such work (Kennicutt et al., 2016) and to make sure that its implications reach policy-makers (CEP, 2017).

3.5.5.5 Role of Informal Actors

3.5.5.5.1 Arctic

Informal actors in the Arctic can influence the decision-making process and policy shaping formulation processes at the AC with regards to legal acts of hard and soft-law at different levels of governance. They are currently providing the shift in AC governance toward more cooperation with distinct actors groups and enhance co-production of knowledge. Informal actors in the Arctic can help to identify and understand the formation of new forums such as the AEC and the Arctic Offshore Regulation Forum (AORF). Recently, the two observers at the AC, the WWF (World Wildlife Fund) and the CCI (Circumpolar Conservation Union) have played an important role in raising awareness and producing scientific reports in the offshore sector in the Emergency, Prevention, Preparedness and Response Working Group (WG EPPR) of the AC and the related Task Force on Arctic Marine Pollution Prevention (TFOPP) (Keil and Knecht, 2017). The WWF recently played an important role in WGs meetings to promote the protection of Arctic biodiversity and the sustainable use of natural resources to influence national policies. However, few studies have concentrated on the role of informal actors and their role in Arctic governance is still at the nascent stage (Duyck, 2011; Makki, 2012; Keil and Knecht, 2017)(*low confidence*).

3.5.5.5.2 Antarctic

Within the ATS several non-state actors play a major role in providing advice on and influencing the governance of Antarctica and the Southern Ocean. Most prominent among these actors at the Antarctic Treaty Consultative Meetings are formal Observers such as the SCAR, and invited experts such as the IAATO and Antarctic and Southern Ocean Coalition (ASOC). At the meetings of the Convention on the Conservation of Antarctic Marine Living Resources, invited observers include organisations such as ASOC, IAATO and SCAR, and representatives of industry such as the Association of Responsible Krill harvesting companies (ARK). SCAR's 2009 report on Antarctic Climate Change and the Environment (ACCE) (Turner et al., 2009) precipitated an Antarctic Treaty Meeting of Experts on Climate Change in 2010 (Antarctic Treaty Meeting of Experts, 2010). The outcomes of the meeting led the Antarctic Treaty's Committee for Environmental Protection (CEP) to develop a Climate Change Response Work Programme, which is now overseen by a formal Subsidiary Group on Climate Change Response (ATCM, 2017). SCAR's ACCE Report updates continue to inform decisions taken by the ATCPs.

Resilience assessments undertaken by stakeholders can facilitate the development of management and governance strategies for assessing risk and coping with change, and through greater awareness and preparedness, thus improves a system's (and community's) capacity to respond. To be effective, the assessment must be meaningful to stakeholders and those who make decisions affecting the system. Thus, it should be integrative, participatory, and aimed at supporting social learning. The Resilience Assessment Framework (Resilience Alliance, 2010) is one method that engages practitioners and researchers in understanding how integrated social-ecological systems change, in order to inform management practices. One activity involves the identification of possible future thresholds and state changes. Evaluating the adaptive and transformative capacity of stakeholders intentionally is complementary with other tools and strategies, with a stronger focus on planning.

3.5.6 Practices for Building Resilient Pathways in Polar Regions

Below we describe select practices currently applied in Polar Regions with potential to contribute to the development of resilience climate pathways. In many cases, the practices described are in early stages of development. Others are well developed. But all, have shown sufficient utility to merit further application with experimentation *(high confidence).*

3.5.6.1 Knowledge Co-production and Integration

The challenges of a changing climate require a new paradigm in knowledge production that moves beyond single disciplinary investigations to an integration and insights from a diversity perspectives (disciplinary and cultural) (Armitage et al., 2011; Petrov et al., 2016; Rycroft-Malone et al., 2016; Berkes, 2017; Miller and Wyborn, 2018; Robards et al., 2018) with production including consideration and integration of a diversity of knowledge systems (i.e., remote sensing analyses, indigenous and local knowledge, conventional biophysical field research, ethnography, simulation modelling, etc.) that contribute both to greater understanding and to improved adaptation. Working in teams, across disciplines and or culture groups, is a necessary part of the process. The process of knowledge co-production includes monitoring (i.e., identification of critical variables/ indicators; data collection; data archiving), improvement of understanding to identify causality or patterns, and communication and review of findings with others to inform decisions (e.g., See US SEARCH Program www.searcharcticscience.org).

An important activity area in Arctic knowledge co-production is in the development and implementation of innovative community-based monitoring (CBM) initiatives (Lovecraft et al., 2013; Johnson et al., 2015a; Johnson et al., 2015b; Tomaselli et al., 2018). One example of innovation in CBM is The Local Environmental Observer (LEO) Network (https://anthc.org/what-we-do/community-environment-and-health/leo-network/) which is using mobile phone technology and the internet to collect, communicate, and discuss unusual observations. Experience shows that CBM provides an opportunity for local and traditional knowledge to interact and in some cases be integrated with other knowledge systems, and potentially contribute to the policy process. While having great potential, however, executing CBM for knowledge co-production is labor intensive, requiring sufficient financial and human resources, on-going refinement of practice, boundary organizations, and strong trust relationships with parties external to community and a long-term commitment by funders and participants to the program (Robards et al., 2018). Several reviews and international initiatives have sought to advance CBM practice and support to nascent stage programs with guidelines, such as ELOKA, a circumarctic initiative to address and disseminate methods (Pulsifer et al. (2012), and https://eloka-arctic.org) , CAFF (See https://www.caff.is/community-based-monitoring/community-based-monitoring-publications), and the Arctic Observing Network (Lee et al., 2015). And as with all knowledge production process, power relationship (who decides, who is ignored, who benefits) underpin collaboration in Arctic and Antarctic science. One possible outcome, therefore, is that CBM can function in a separate sphere from more conventional science efforts, with little interaction. CBM in the Arctic is one practice of a growing field of cross-cultural, interdisciplinary programs whose potential and utility in supporting human adaptation are not yet fully realized. Focused on-going efforts are needed to develop the field of practice *(high confidence).*

3.5.6.2 Linking Science with Decision Making

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

There is a growing expectation in Polar Regions for a more deliberate strategy of co-producing science and policy in an iterative process of regular interactions among scientists, their managers, officials, and stakeholders. This redefined “actionable science” can better support decisions by creating more rigorous and accessible products while shaping a narrative that instills public confidence (Beier et al., 2015; Fleming and Pyenson, 2017). modeling, structured decision making systems, and visualization, and decision theaters are a few methods recently employed and being developed to link science and policy in the Polar Regions (Schartmüller et al., 2015; Kofinas et al., 2016; Garrett et al., 2017; Holst-Andersen et al., 2017; Camus and

Smit, 2018). A key adjustment to business as usual, however, involves willingness to provide active decision-support *services,* more often than mere decision-support *products* (Beier et al., 2015). In short, circumstances call for a new breed of Polar scientist who not only understands policy considerations, but also engages in adaptive co-management stages of policy formulation. Polar scientists can do much more to make their work widely available for use, including: enhanced data collaboration at every scale, more strategic social engagement, and explicit creation of consensus documents that provide interpretive guidance about research implications and alternative choices (Gewin, 2014). In many cases, successful efforts in linking science with policy follow from basic communication and personal relationships of trust *(high confidence.)*

3.5.6.3 Scenario Analysis

Given the need to assess future risk and significant challenges of responding to Polar climate change in conditions of uncertainty, methods of exploring plausible, likely and desirable futures with stakeholders, scientists, and policy makers are needed (Resilience Alliance, 2010; Arctic Resilience Report, 2016; Flynn et al., 2018). Participatory scenario analysis in the Arctic has been implemented using many approaches for a variety of problem areas. The Canadian Department of National Defence used scenario analysis to study the national security issues of an ice-free Arctic. Another is in the Barents region, where scenario workshops have included local and regional actors from public agencies, organizations and the private sector in three different locations (Pajala, Sweden; Kirovsk, Russia; and Bodø, Norway). Exploring possible futures was undertaking together with reindeer-herding youth across the Eurasian Arctic (Oort et al., 2015; Nilsson et al., 2017). The innovative Scenarios Network for Alaska and Arctic Planning (SNAP) downscales GCMs to local regions, communicates data and outputs in user friendly formats, and engages of stakeholders through partnership programs (See https://www.snap.uaf.edu) (Flynn et al., 2018) review participatory scenario analysis found that less than half of those studies identified incorporated climate projections. For example the The Oil Development Scenarios Project of the North Slope Science Initiative of Alaska, which used maps in a participatory analysis process and led to the identification of research needs, did not consider climate as a driver of change. Flynn et al also found that most studies utilize a forecasting approach, with those utilizing a backcasting approach having higher local participation, and studies that integrated different knowledge systems, cultural factors may impact the utility and acceptance of the approach. Clearly, the participatory approach to scenario analysis has potential to enhancing knowledge co-production, and facilitating proactive adaptation *(medium confidence).*

3.5.6.4 Assessing Perceived Risk in Addition to Probability-based Risk

Measureable risk provides a science-based indication of the vulnerabilities to climate change, such as the calculation of the likelihood of a fuel spill due to the increase in shipping in the Polar Regions (Gascard et al., 2017) or the probable cost of damaged infrastructure due to thawing permafrost (Melvin et al., 2017b) . Human perceptions of risk may or may not align with science-based measurable risk and analysis of both types of risk can provide insight into vulnerabilities, while revealing reasons for the lack of human awareness and action or a possible oversight in other assessing risks (Blair et al., 2014; Blair, 2017)*(medium confidence).*

3.5.6.5 Self Assessments of Social-ecological Resilience to Build Adaptive and Transformative Capacity

Resilience assessments undertaken by stakeholders can facilitate the development of management and governance strategies for assessing risk and coping with change, and through greater awareness and preparedness, thus improves a system's (and community's) capacity to respond. To be effective, the assessment must be meaningful to stakeholders and those who make decisions affecting the system. Thus, it should be integrative, participatory, and aimed at supporting social learning. The Resilience Assessment Framework (Resilience Alliance, 2010) is one method that engages practitioners and researchers in understanding how integrated social-ecological systems change, in order to inform management practices. One activity involves the identification of possible future thresholds and state changes. Evaluating the adaptive and transformative capacity of stakeholders intentionally is complementary with other tools and strategies, with a stronger focus on planning.

3.5.6.6 Resilience-based Ecosystem Stewardship

Resilience-based ecosystem stewardship, by definition, differs from conventional resource management or ecosystem management, while retaining many of the principles of those two paradigms (Chapin et al., 2009; Chapin et al., 2010a)(See Table 3.6) In the Polar Regions, stewardship of resources requires a focus on trajectories of change, implying maintaining ecosystems in a state of equilibrium is not possible (Biggs et al., 2012; Arctic Resilience Report, 2016). Several responses consistent with stewardship approach are practised to reduce impacts and risks to species, habitats and ecosystems in support of biodiversity resilience. The first line implements the tools of biodiversity conservation. Often expressed to protect the intrinsic values of biodiversity, they are increasingly understood as also supporting sustainable use of the environment (Ban et al., 2014), secure options for livelihoods (Salafsky and Wollenberg, 2000), and facilitating biodiversity adaptation in a changing environment (Mawdsley et al., 2009). In particular, networks of protected areas (vs isolated protected areas) are conceptualised (McLeod et al., 2009), planned (Solovyev et al., 2017) and implemented (Juvonen and Kuhmonen, 2013) to protect ecologically connected tracts of representative habitats, and biologically and ecologically significant features. While individual protected areas may prove problematic in a rapidly changing ecosystem, protected area networks that combine both spatially rigid and spatially flexible regimes and also include climate refugia operate in support of ecological resilience to climate change by maintaining genetic connectively and flows, reducing direct pressures on biodiversity, and thus, giving biological communities, populations, and ecosystems the space to adapt (Nyström and Folke, 2001; Hope et al., 2013). The second area of response seeks to maintain a continued flow of ecosystem services that meet human needs and use the recognition of the benefits that humans derive from these services to provide incentives for preserving biodiversity while ensuring options for sustainable use of resources and economic development (Guerry et al., 2015). Incorporating Arctic ecosystem services into policies and governance practices and capturing them in decision-making processes strengthens the resilience of Arctic social-ecological systems to rapid changes and is a key method for the integration of environmental, economic, and social policies (CAFF, 2015). Currently however, using the approach in the Arctic is held back by a lack of comprehensive recognition of the wide range of benefits people receive from Arctic ecosystems and by a lack of planning and management tools that can demonstrate these benefits in decision-making processes (CAFF, 2015). A third line focuses on shaping pathways of social-ecological change with the goal to foster a more sustainable future for species, habitats, ecosystems, communities, and society (Chapin III et al., 2015). Such processes require engagement of a diversity of stakeholders to define problems, solutions and actions, thereby providing a basis for fostering social learning as a part of the resource management process (as described in the strategies outlined above) (Knapp et al., 2014) *(high confidence).*

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

Table 3.6. Differences between conservation approaches focused on species, landscapes, and the mutual wellbeing of people and nature. (Chapin III et al., 2015)

Characteristic	Species Conservation	Landscape and Seascape Conservation	Stewardship
Reference point	Historic condition	Historic and current condition	Pathways of change
Central goal	Species protection	Conservation of ecosystem structure and function to conserve biodiversity and the habitats that support it	Sustain social-ecological systems and resilience of ecosystem services by fostering diversity

Predominant approach	Maintain species, populations, and habitats	Integrated management of human activities in landscapes and seascapes	Manage stabilizing and amplifying feedbacks
Role of protected areas	Habitat that is relatively safe from direct human impacts	Part of the habitat mosaic that interacts with unprotected habitat	Part of a complex social-ecological system that supports conservation and interacts with other societal goals
Role of uncertainty	Reduce uncertainty before taking action	Reduce uncertainty yet act in its presence	Embrace uncertainty: Maximize flexibility to adapt to an uncertain future
Role of resource manager(s)	Decision maker who sets course for sustainable management of species, populations, and habitats	Decision maker who sets course for sustainable management of landscapes, seascapes, and their components	Coordinated facilitators at multiple scales who engage stakeholder groups to respond to and shape social-ecological change and nurture resilience
Response to disturbance	Minimize disturbance probability and impacts	Minimize disturbance probability and impacts	Disturbance cycles used to provide windows of opportunity
Resources of primary concern	Species, populations, and habitats	Species, populations, and landscapes or seascapes	Biodiversity, human wellbeing, and adaptive capacity

3.5.7 Conclusion

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

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Appendix 3.A: Supplementary Material

The material contained in the appendix is planned as online supporting material to the final chapter/report.

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

3.A.1.1 Northern Hemispheric Climate Modes

While the Northern Hemisphere atmospheric wind motion is primarily a north-south wavy jet stream pattern that can consist of propagating multiple waves, another approach to specifying atmospheric circulation variability is through atmospheric modes that can capture historical variance patterns, often through mathematical methods such as Empirical Orthogonal Functions (EOFs).

The first pattern is the Arctic Oscillation (AO) or Annular mode that in its positive sign has zonal symmetric flow centered on the North Pole. In its negative sign this pattern breaks down into a weaker and move wavy circulation pattern.

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

Other patterns of interest is the Arctic Dipole (AD), which is the third hemispheric pattern. In contrast to the AO that is circular around a latitudinal circle, the AD has flow across the central Arctic with high and low pressures on either side (Asia and North America). The North Atlantic Oscillation (NAO) appears to be an Atlantic extension of the AO with a positive sign for lower pressure near Iceland.

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

3.A.1.2 Ice-Albedo Feedback and Polar Amplification

Early studies hypothesized that the impacts of global warming would first be manifested in the polar regions as increases in air temperatures would lead to reductions in snow and ice, allowing more of the sun's energy to be absorbed by the surface, fostering more melt (Manabe and Stouffer, 1980). As the sea ice in the Arctic has retreated, the expanding open water areas in summer absorb much of the Sun's energy, warming the ocean mixed layer. Before sea ice can reform in winter, the ocean must release the heat gained in summer back to the atmosphere. This leads to strong low-level atmospheric warming, which in part explains why the Arctic has warmed about twice as fast as the mid-latitudes (Overland et al., 2017a) (see Box 3.1). Furthermore, increased exchanges of latent heat flux from the ocean to the atmosphere has led to increased atmospheric water vapor (Serreze et al., 2012). The sea ice-albedo feedback (Perovich et al., 2008), has thus been largely implicated in the observed Arctic amplification of warming trends (Serreze et al., 2009; Screen and Simmonds, 2010) (*very high confidence*).

While the sea ice-albedo feedback links reductions in summer sea ice cover to warming in autumn and winter, there is emerging evidence of increased warm, moist air intrusions in both winter and spring (Kapsch et al., 2015; Boisvert et al., 2016; Cullather et al., 2016; Mortin et al., 2016; Graham et al., 2017). Tropical convection may play an important role by exciting moisture intrusion events on inter-decadal time scales (Lee et al., 2011). Furthermore, intra-seasonal tropical convection may influence daily Arctic surface temperatures in both summer and winter (Yoo et al., 2012a; Yoo et al., 2012b; Henderson et al., 2014).

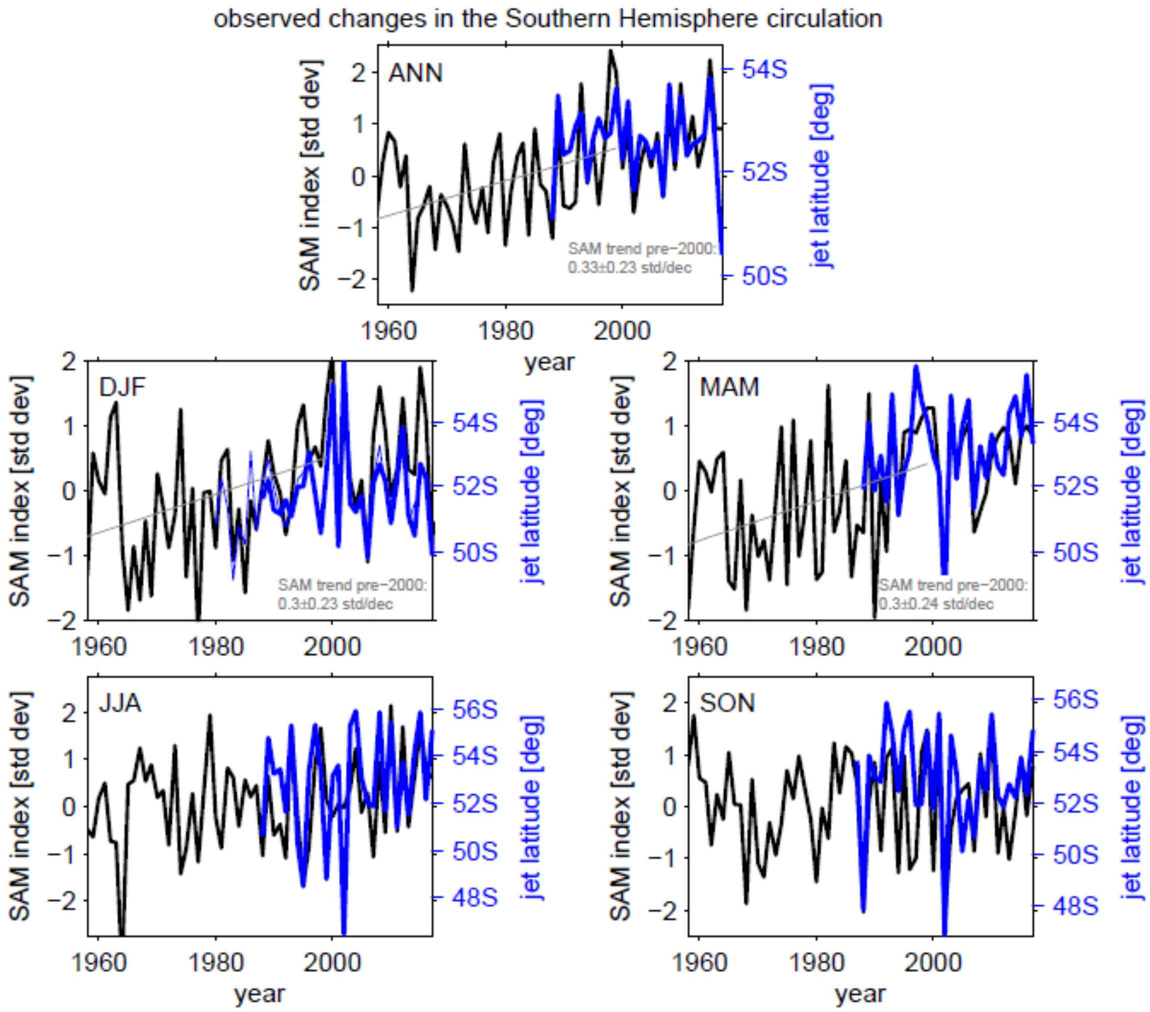
These events lead to increased downwelling longwave radiation from a warmer free troposphere as well as a change in optical depth from increased atmospheric moisture. Modeling studies have shown that Arctic Amplification occurs in the absence of the sea ice-albedo feedback (Alexeev et al., 2005), and the largest contributor to Arctic Amplification is increased downwelling longwave radiation rather than the ice-albedo feedback (Pithan and Mauritsen, 2014). It is important to clearly differentiate the contributions from local forcing (i.e., ice-albedo feedback, increased atmospheric water vapor and cloud cover) from remote forcing (i.e., changes in atmospheric circulation).

3.A.1.3 Southern Hemispheric Climate Modes

The Southern Hemisphere extratropical circulation is much more zonally symmetric than the Northern Hemisphere, with a strong belt of westerly winds encircling the Antarctic continent. In winter and spring these winds exhibit more zonal asymmetries, expressed by the zonal wave 3 and Pacific South American patterns (Irving and Simmonds, 2015). Understanding decadal variability, such as the Pacific Decadal Oscillation/Interdecadal Pacific Oscillation's (PDO/IPO) impact on these modes is hampered by the shortness of the observational record, with limited station data available poleward of 40S (Marshall, 2003). Observed changes in the Southern Hemisphere extratropical atmospheric circulation is primarily indicated by the Southern Annular Mode (SAM), the leading mode of extratropical variability in sea level pressure or geopotential heights which is related to the latitudinal position and strength of the mid-latitude eddy-driven jet. The SAM has a strong influence on the weather and climate of SH polar regions as well as southern Australia, New Zealand, southern South America and South Africa (see review article by Thompson et al. (2011)).

Numerous studies have attributed the significant poleward shift and strengthening of the SAM over the past 30-50 years to anthropogenic forcing, in particular stratospheric ozone depletion and increasing greenhouse gases (Appendix 3.A, Figure 1). Though the exact mechanisms by which these forcings impact the circulation is unclear, they both act to enhance the meridional temperature gradient which leads to a poleward shift in the mid-latitude jet. There is *medium confidence* that ozone depletion is the dominant driver of recent austral summer changes in the Southern Hemisphere circulation during the period of maximum ozone depletion from the late 1970s to late 1990s (Arblaster et al., 2014; Waugh et al., 2015). In the years following, Waugh et al. (2015) and other studies argue for a strong impact of tropical Pacific sea surface temperatures in driving positive SAM trends (Schneider et al., 2015a; Clem et al., 2017a).
Zonal wave 3 (ZW3) describes the asymmetric part of the generally strongly zonally symmetric circulation in the SH extratropics and has been shown to impact the SH surface climate, blocking, sea-ice extent and the strength of the Amundsen Sea Low (Turner et al., 2017a; Schlosser et al., 2018). It has its strongest amplitude in SH winter and is more prominent during phases of negative SAM (Irving and Simmonds, 2015). No significant trends in the amplitude or phase of zonal wave 3 over the satellite era have been found (Turner et al., 2017c).

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



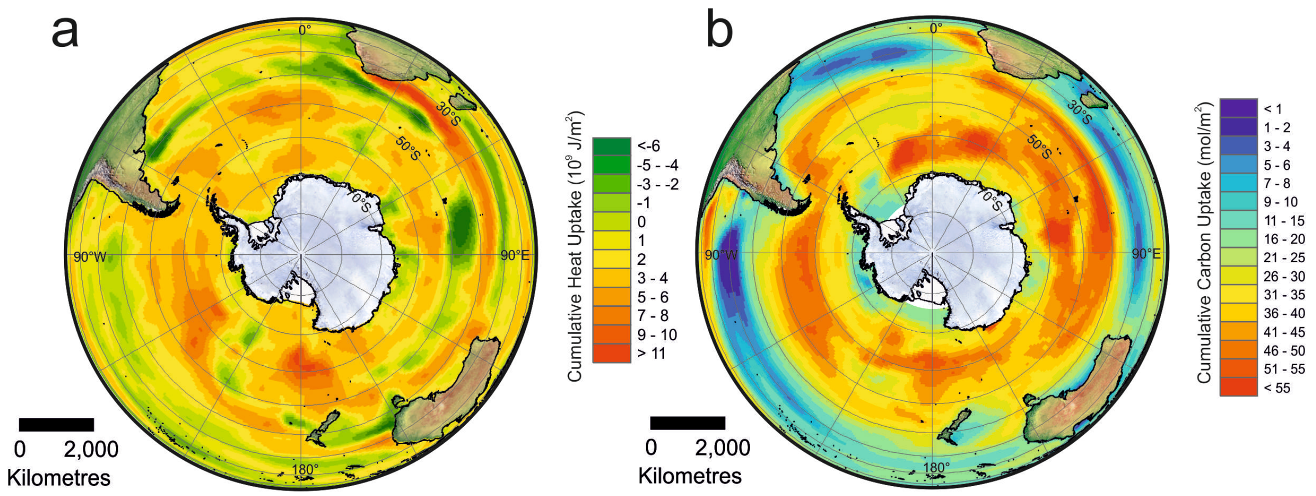
Appendix 3.A, Figure 1: SAM index (black) and mid-latitude jet positions (blue) time series for the four seasons and annual mean. The SAM index is normalized by its standard deviation, and is defined as in Marshall, 2003. The jet position is based on the maximum of CCMP satellite-based surface wind speed (Atlas et al. 2011; available for download at http://www.remss.com/measurements/ccmp.html) which starts in 1987. For DJF, we also show jet position from the MERRA reanalysis. A linear trend line of the SAM changes before 2000 is shown when statistically significant, and the slope of the best fit line and its corresponding 95% uncertainty bounds are shown

3.A.2 Changes in Polar Ice Sheets and Glaciers: Ocean, Sea Level and Ecosystem Impacts

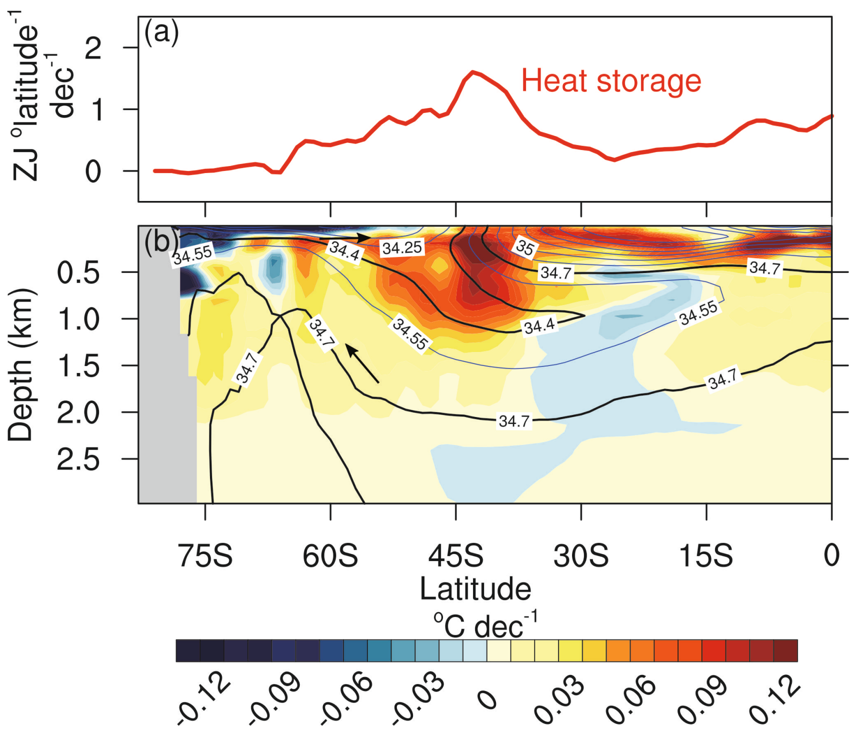
[PLACEHOLDER FOR SECOND ORDER DRAFT]

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

3.A.3.1 Heat and Carbon Uptake by the Southern Ocean

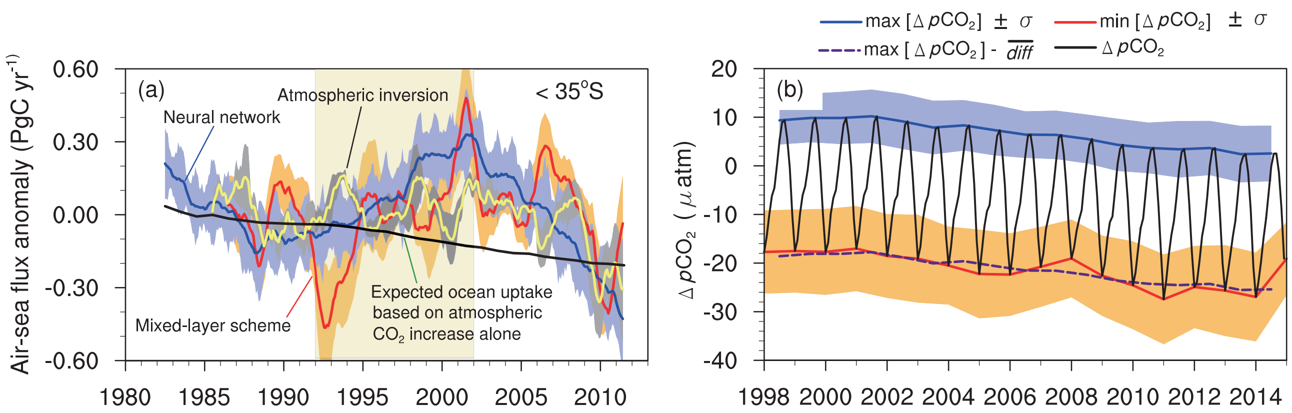


Appendix 3.A, Figure 2: CMIP5 multimodel mean changes in depth-integrated oceanic heat (a) and anthropogenic carbon (b) between 1870 (represented by mean of period 1861-80) and 1995 (represented by mean of period 1986-2005). Data from Frölicher et al. (2015).



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

3.A.3.2 Decadal Variability in the Southern Ocean Air-sea Flux of CO_2



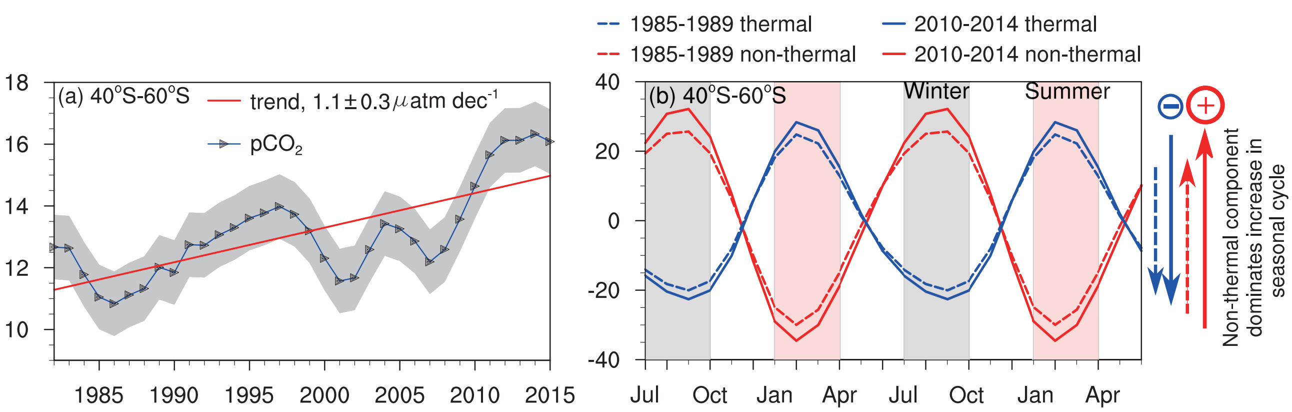
Appendix 3.1, Figure 4: (a) Decadal variability in the Southern Ocean air-sea CO2 flux anomaly (adapted from Landschützer et al. (2015)). Curves contrast the decadal model reconstruction (1982-2012) of CO2 air-sea flux anomalies from observations and neural network (blue) against a second empirical method (Rodenbeck et al., 2014) and a model-based steady-state linear trend of an increasing CO2 sink. (b) The interannual variability of the seasonal cycle of ΔpCO2 showing that the decadal trend (1998-2012) is strongly associated with trends in winter peaks of ΔpCO2, whereas the summer minima have stronger interannual modes (adapted from Gregor et al. (2017b))

3.A.3.3 Variability and Trends in DIC Buffer Factor (γ)

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

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

This has been further investigated using observation-based CO_2 products (Landschützer et al., 2018). Using the data product that spans 34 years (1982-2015) the study confirms the model predictions that there already exists a an observable trend in the increase of the mean seasonal amplitude of the seasonal cycle of pCO_2 of 1.1 ± 0.3µatm per decade in the Southern Ocean (Landschützer et al., 2018)(Appendix 3.A, Figure 5a). It also shows that this mean trend is the net effect of opposing forcing from biogeochemical (non-thermal) (2.9 ± 0.7) and thermal (-2.1 ± 0.5) (Appendix 3.A, Figure 3.A.3.5b). Overall, these changes to the characteristics of the seasonal cycle of biogeochemistry and CO_2 because of the trends in reduced buffering will become dominant drivers of the long-term trend of the fluxes and storage of anthropogenic CO_2 in the Southern Ocean (McNeil and Sasse, 2016).



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

3.A.3.4 Decadal Changes in Southern Ocean Carbon Storage ates

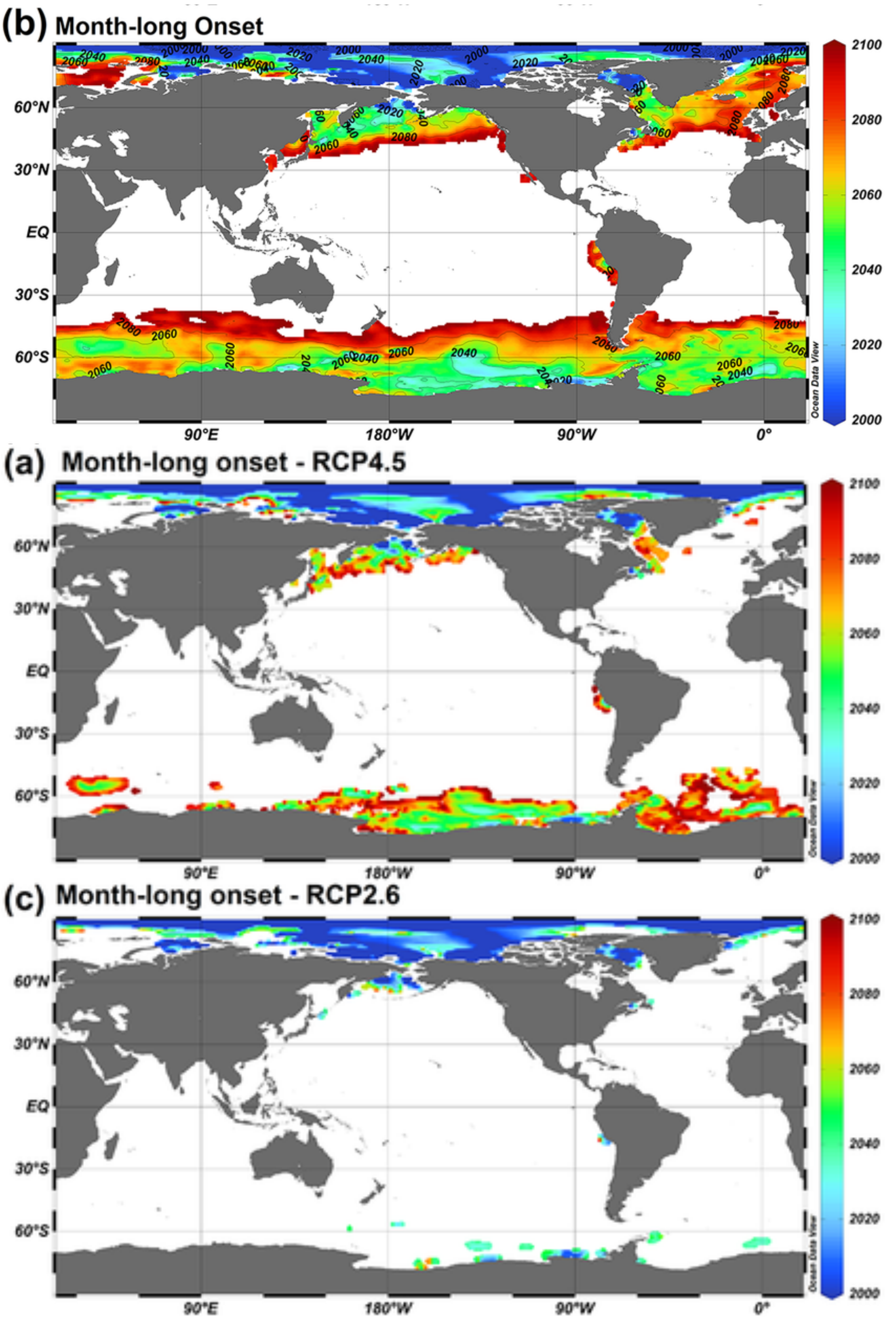
Decadal changes in the modelled net carbon and observed anthropogenic carbon (C_{ANT}) storage rates a may be linked to the decadal phases of the Upper Ocean Overturning Circulation (UOOC) (Appendix 3.A, Table 1). The net carbon storage is largely influenced by changes in the outgassing flux as a response to the intensification or weakening of the upwelling of Upper Circumpolar Deep Water (UCDW). This has the potential to explain why storage increases when UOOC weakens and outgassing is reduced (DeVries et al., 2017). The magnitude of the carbon storage variability is an indication of the sensitivity of the system to small wind-driven adjustments in the UOOC. In contrast, C_{ANT} has maximum storage during high UOOC phases probably due to its sensitivity to the increased rate of subduction of Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) (Tanhua et al., 2017).

Appendix 3.A, Table 1: Compares the phasing and magnitude of the decadal variability in net carbon and anthropogenic carbon storage in the Southern Ocean (DeVries et al., 2017; Tanhua et al., 2017).

Decade	DeVries (2017)		Tanhua (2016)	
	Net storage CO_2	Explanation	C_{ANT} Storage Rates	Explanation
1980s	High - 0.53$PgCy^{-1}$	Slow UOOC Outgassing reduced & storage increased	1984-1990 440kmoly-$^{1}m^{-1}$	Lower storage in SAMW
1990s	Low - 0.20$PgCy^{-1}$	Faster UOOC Outgassing: increased storage reduced	1984-2005 1142$kmoly^{-1}m^{-1}$	High Storage in SAMW
2000s	High - 0.61$PgCy^{-1}$	Slow UOOC Outgassing: decreased storage increased	2005-2012 -752$kmoly^{-1}m^{-1}$	Lower Storage in AAIW

Appendix 3.A, Table 2: Timing of the onset of monthly and annual mean undersaturation in the Southern Ocean under different AR5 defined emission scenarios. It highlights the impact tipping point between RCP2.6 and RCP4.5/8.5.

Scenario	**Onset of Month USat**	**Onset of annual USat**	**% Impact area reduction rel RCP8.5**
RCP 8.5	2030 – 2080	+ 10 - 20	-
RCP 4.5	2064 ± 17	+ 10 – 20	
RCP 2.6	2033 ± 15	- None	99.8%



Apendix 3.A, Figure 6: The comparative spatial extent of seasonal carbonate undersaturation by 2100 in the Southern Ocean under contrasting emission scenarios RCP8.5/4.5/2.6 (Sasse et al., 2015). [to be redrafted from original data for second order draft]

3.A.3.5 Climate Change Impacts on Arctic Kelp Forests

In the Arctic, biodiversity of macroalgae and biomass of kelps and associated fauna have considerably increased in the intertidal to shallow subtidal over the last two decades, causing changes in the food web structure and functionality. This is mostly attributed to the reduced physical impact by ice-scouring and increased light availability as a consequence of warming and concomitant fast-ice retreat (Kortsch et al., 2012; Bartsch et al., 2016; Paar et al., 2016) (*medium confidence*). Increase of summer seawater temperatures up to 10°C (IPCC 2100 scenario) will not be detrimental for Arctic kelp species. A further seawater temperature increase above 10°C which is only expected under extreme warming scenarios will definitely suppress the abundance, growth and productivity of Arctic endemic *Laminaria solidungula* and sub-Arctic *Alaria esculenta* but not of cold-temperate to Arctic *L. digitata* and *S. latissima* (Dieck, 1992; Gordillo et al., 2016; Roleda, 2016; Zacher et al., 2016) (*high confidence*). In total, these data support projections that kelp and macroalgal production will increase in the future Arctic (e.g., Krause-Jensen and Duarte 2014). This will become more pronounced when rocky substrates hidden in current permafrost areas (Lantuit et al. 2012) will be readily colonized by kelp and other macroalgae when getting ice-free as has been verified for Antarctica (Liliana Quartino et al., 2013; Campana et al., 2017)(*high confidence*).

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

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

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

[PLACEHOLDER FOR SECOND ORDER DRAFT]

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

[PLACEHOLDER FOR SECOND ORDER DRAFT]