AMBIO (2011) 40:75–86
DOI 10.1007/s13280-011-0215-8

Feedbacks and Interactions: From the Arctic Cryosphere to the Climate System

Terry V. Callaghan, Margareta Johansson, Jeff Key, Terry Prowse, Maria Ananicheva, Alexander Klepikov

Abstract Changes in the Arctic's climate are a result of complex interactions between the cryosphere, atmosphere, ocean, and biosphere. More feedbacks from the cryosphere to climate warming are positive and result in further warming than are negative, resulting in a reduced rate of warming or cooling. Feedbacks operate at different spatial scales; many, such as those operating through albedo and evapotranspiration, will have significant local effects that together could result in global impacts. Some processes, such as changes in carbon dioxide (CO_2) emissions, are likely to have very small global effects but uncertainty is high whereas others, such as subsea methane (CH_4) emissions, could have large global effects. Some cryospheric processes in the Arctic have teleconnections with other regions and major changes in the cryosphere have been largely a result of large-scale processes, particularly atmospheric and oceanic circulation. With continued climate warming it is highly likely that the cryospheric components will play an increasingly important climatic role. However, the net effect of all the feedbacks is difficult to assess because of the variability in spatial and temporal scales over which they operate. Furthermore, general circulation models (GCMs) do not include all major feedbacks while those included may not be accurately parameterized. The lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs.

Keywords Feedbacks · Ocean circulation · Atmospheric circulation · Cryosphere · Teleconnections

Electronic supplementary material The online version of this article (doi:10.1007/s13280-011-0215-8) contains supplementary material, which is available to authorized users.

INTRODUCTION

Feedback processes are responses to a driving mechanism that subsequently accelerate (positive feedback) or retard (negative feedback) the original driving process. At the end of the nineteenth century, Arrhenius (Bolin 2007) described the classic feedback whereby increased air temperature leads to an increase in the amount of water vapor in the atmosphere, which in turn leads to additional atmospheric warming. More recent studies detailed more feedbacks (e.g., Francis et al. 2009). However, a new comprehensive assessment of changes in the cryosphere and their consequences (SWIPA—snow, water, ice, permafrost in the Arctic: AMAP 2011) has enabled a new assessment of feedbacks from the cryosphere to the atmosphere to be made (Callaghan et al. 2011a). This article distills the key findings from the feedbacks portion of the SWIPA report.

There are many feedbacks from the cryosphere to the climate system: some are direct, but others are complex and indirect. Some of the complexity and variety of interactions between the atmosphere, ocean, and cryosphere are illustrated in Fig. 1. Further complexity arises owing to scaling issues: individual feedbacks operate over different time scales and their effects can vary from local to global spatial scales. Furthermore, some feedbacks to climate warming are negative and result in reduced rates of climate warming or climate cooling whereas others are positive and lead to warming. An overall assessment of the net effect of many different potential feedbacks on climate change has yet to be achieved. This article summarizes the feedbacks associated with a changing Arctic cryosphere presented in AMAP (2011), describes other feedbacks and interactions that span the various cryospheric components, and provides a preliminary assessment of their relative magnitudes. However, the calculation of the net effects of all feedbacks requires complex modeling that remains a priority for future research.

FEEDBACKS

Greenhouse Gases

Carbon Dioxide (Land, Freshwater, and Marine)

Carbon dioxide is a greenhouse gas with a radiative forcing of 1.66 Wm^{-2}. The adjustment time of CO_2 in the atmosphere is ~100 years (Forster et al. 2007). It is exchanged between the atmosphere and the biosphere through the processes of photosynthetic capture by green plants, short-term autotrophic respiration by plants, and long-term heterotrophic respiration of dead plant material by microbes. A mismatch, particularly in wet areas, between the rate of fixation of atmospheric CO_2 and its release from the biosphere has led to a net accumulation of carbon in the Arctic with considerable amounts preserved in permafrost (AMAP 2011). Around 44% of the world's near-surface labile soil carbon is found in Arctic soils (McGuire et al. 2009; AMAP 2011) and permafrost together with the overlying active layer contain approximately twice as much carbon as the global atmosphere (McGuire et al.

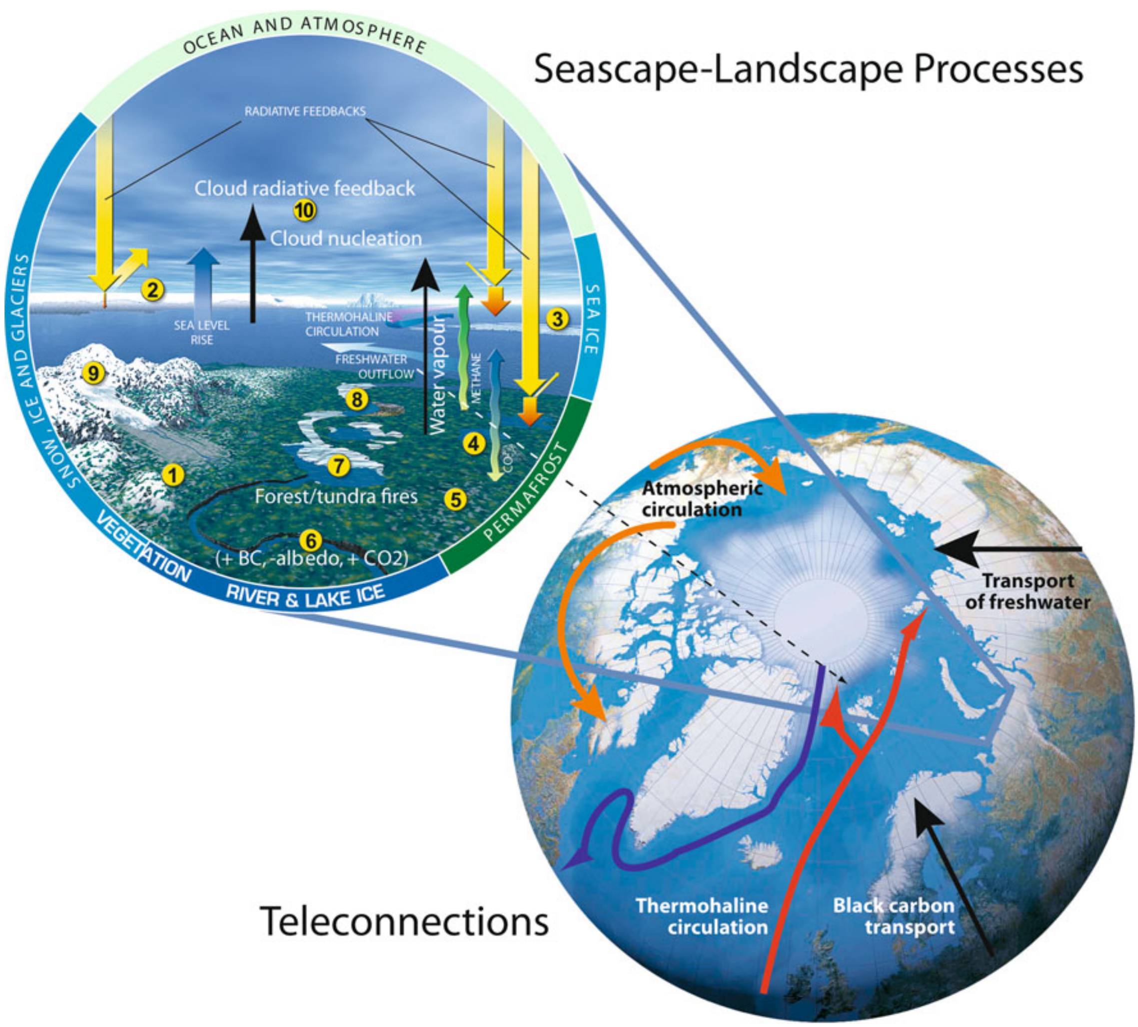

Fig. 1 Local, regional, and global feedbacks and processes related to changes in the Arctic cryosphere. The *numbered yellow circles* refer to impacts and processes: (1) melting and retreating snow cover increases radiation absorption, a radiative feedback. (2) Melting of large ice sheets contributes to sea-level rise and the freshwater flux with potential effects on thermohaline circulation and global climate. (3) Retreating sea ice contributes to increased radiative absorption (ice-albedo feedback) and heat and moisture fluxes to the atmosphere, which impact cloud cover. (4) As permafrost degrades, CH_4 production increases. With wetland drying, CO_2 emissions increase, and the atmosphere warms over time. (5) Thawing permafrost changes geomorphic/geochemical processes and fluxes. (6) Increasing precipitation plus melting snow and ice increases river flow and changes the freshwater flux. (7) Shrinking lake-ice cover has ecological impacts generally leading to greater productivity but negatively impacting surface transport. (8) Changes in the magnitude and timing of snowmelt runoff and river-ice processes have both positive and negative impacts. (9) Retreating glaciers initially increase runoff but lower flows eventually result as ice masses diminish. (10) Changes in cloud cover affect the surface radiation budget, which impacts sea ice, in turn affecting cloud cover (cloud-radiation feedback). Source: after Prowse (2009)

2009; Tarnocai et al. 2009). This carbon store is sensitive to climate warming.

The major cryospheric controls on carbon *capture* from the atmosphere are the duration and timing of the snow-free period. Euskirchen et al. (2006) estimated increased carbon drawdown at about 9.5 g C m^{-2} $year^{-1}$ for each day that the growing season was projected to increase between 2001 and 2100. Recent analyses from optical satellite imagery of changes in snow-cover extent show that decreases are mainly in the spring period (18%) over the 1966–2008 period, thereby extending the snow-free season (AMAP 2011; Callaghan et al. 2011b [this issue]).

The major cryospheric controls on carbon *release* from the soil are thawing permafrost (AMAP 2011) and soil warming during and as a result of extreme events such as fire (Mack et al. 2011). Experimental warming of soils (Dorrepaal et al. 2009) and observations of areas that are experiencing different degrees of warming (e.g., Schuur et al. 2009 and other references in AMAP 2011) show that the Arctic is already losing some of the carbon that has been stored for thousands of years. Between 4.5 and 6.0 kg m^{-2} (or 9.5–13%) of the soil organic matter carbon pool could be lost on a century time scale (Schuur et al. 2009). Currently, widespread permafrost thawing throughout the Arctic has not been reported. However, recent projections of permafrost temperatures show thawing to be widespread throughout the southern regions of the Arctic by 2090 (AMAP 2011).

Overall, models suggest that the Arctic will remain a weak sink of carbon during warming (e.g., McGuire et al. 2009). However, both current and future carbon sink activity can be reversed by short-term disturbances such as forest fires and insect pest outbreaks, and long-term changes. The long-term changes include thawing permafrost and altered hydrology related to permafrost dynamics and changes in snow regime that lead to landscape drying and plant water stress. Also, the models exclude disturbance due to human activities. Although, the future feedback sign and strength for CO_2 in the Arctic remain uncertain, it is likely that century-long processes of negative feedbacks from increased CO_2 sequestration will be interspersed with episodic releases following disturbances.

Freshwaters (lakes, ponds, and rivers) feed back to the atmosphere through processes that occur in their immediate vicinity (e.g., carbon fluxes, albedo, and evaporation) and downstream when the waters enter the Arctic Ocean (see below). As temperatures rise, the number of days of ice cover decreases and more heat is absorbed by the open water, which in turn increases CO_2 capture through the longer duration of primary production activity. However, many northern lakes and ponds are already supersaturated with carbon and are net sources to the atmosphere (Jonsson et al. 2003). Such sources are likely to grow following increases in the supply of organic material from permafrost thawing and carbon transport to the lakes from enhanced terrestrial plant production, but will be determined by the balance between drying and water-logging of the land surface, which is as yet unknown.

The Arctic Ocean could be a significant sink for carbon as sea-ice retreats and the ocean warms. As the ice edge retreats away from the continental shelves during climate warming, more algal blooms could occur, which might lead to more organic material and carbonate in the shells of some plankton species falling to the deeper ocean bed off the continental shelves and accumulating there. However, this negative feedback to climate warming might be moderated by ocean acidification. The pH of the ocean is currently decreasing as more CO_2 is drawn down from the atmosphere: the resulting slightly more acidic conditions affect the species that use calcium to build cell walls causing less calcium to be deposited onto the ocean bed. In contrast, Alekseev and Nagurny (2007) speculated that, overall, the Arctic Basin is more likely to be a source than a sink of CO_2 over an annual cycle.

Methane (Wetlands, Lakes, and Sub-Sea)

In anaerobic soils, ponds and lakes of the Arctic, microbes produce CH_4 rather than CO_2. Methane is 25 times more powerful than CO_2 as a greenhouse gas over a 100-year time frame with a radiative forcing of 0.48 W m^{-2} (Forster et al. 2007), so any increases in CH_4 emissions will be particularly important. Unlike CO_2, CH_4 is difficult to measure over large areas because there are geographical hotspots for CH_4 production (e.g., Walter et al. 2007) and because there are episodic releases of CH_4 (Mastepanov et al. 2008).

In addition, it is difficult to project future releases of CH_4 from land because wetlands on permafrost are drying in some areas (e.g., Smith et al. 2005; AMAP 2011), which would decrease current CH_4 emissions. Conversely, thermokarst pond formation is occurring elsewhere with concomitant increases in CH_4 emissions (e.g., Christensen et al. 2004).

Particularly, large sources of CH_4 occur in former soils that were inundated when sea levels rose after the last ice age. These vast continental shelves of the Arctic such as the Laptev Sea have shown high levels of CH_4 throughout the water column and extending up to 1800 m in the atmosphere (e.g., Shakhova et al. 2010). Overall, a 1% release of CH_4 stored in subsea hydrates in the Arctic would be equivalent to a doubling of atmospheric CO_2 concentration in terms of its radiative effect. Although, the uncertainties of the size of the subsea carbon reserves and their stability are great, the potential risk from increased CH_4 release is sufficiently large to cause concern.

Nitrous Oxide (Tundra Peat Lands)

Very high emissions of the powerful greenhouse gas nitrous oxide (N_2O; radiative forcing of 0.16 W m^{-2}, Forster et al. 2007) have recently been discovered from peat circles in patterned ground of the discontinuous permafrost zone in Russia (Repo et al. 2009). High N_2O emissions of 34 mg N m^{-2} day^{-1} were measured in cores taken from northeastern Greenland and incubated in the laboratory (Elberling et al. 2010). These values are equivalent to daily N_2O emissions from tropical forests on a mean annual basis and emphasise the importance of permafrost N_2O sources, previously thought of as of unimportant.

Water Vapor (Land, Sea, Lakes, and Rivers)

Water vapor is a greenhouse gas with a radiative forcing of 0.07 W m^{-2} (stratospheric water vapor from CH_4; Forster et al. 2007). In the Arctic, it is transported from the surface to the atmosphere via evaporation from open water and wet surfaces, evapotranspiration from vegetation, and sublimation from snow and ice. The feedback effects of the processes generating water vapor are complex in that evaporation leads to local cooling, but the increased atmospheric concentrations of water vapor can lead to warming over a wider area because of mixing and transport of water vapor in the atmosphere (positive feedback). It is further complicated by the possibility of increased atmospheric water vapor enhancing cloud formation, which can have either a warming or cooling effect on the surface depending on the cloud height and time of year (Rouse et al. 1997, also discussed below).

Evapotranspiration is expected to increase in those areas of the Arctic where plant production increases. Projections of increased forest growth and extent in the Barents region of the Russian Arctic show that by 2080 summer temperatures will have decreased by 1.5°C due to evaporative cooling (Gottel et al. 2008). However, other feedbacks are also associated with changes in vegetation, such as albedo.

Ozone and Bromine (Marine)

Releases of bromine gas species from the ocean, possibly via sea-salt aerosol production from snow lying on sea ice during blowing snow events, can result in sudden ozone depletion in the lower troposphere during spring. A reduction in Arctic sea ice would reduce the importance of this process in the chemistry of the Arctic atmosphere. In chemistry model simulations, Voulgarakis et al. (2009) found large spring ozone increases (up to 50–60%) over the Arctic, due mainly to a reduction in the impact of bromine chemistry, caused by sea-ice retreat. Tropospheric ozone has a relatively small radiative impact (warming), although the effect is greater over bright surfaces (Shindell et al. 2006). With a declining ice cover potentially giving rise to an increase in tropospheric ozone, the feedback would be positive. Scinocca et al. (2009) found that the stratospheric response in springtime polar cooling is dynamical rather than radiative in origin. In the model simulations, the response lags the onset of the sea-ice loss by about a decade. It is associated with an enhanced weakening of the North Atlantic Meridional Overturning Circulation (i.e., the strength of the thermohaline circulation in the North Atlantic).

Transfer of Heat and Energy Between the Cryosphere, Atmosphere, and Ocean

Albedo

A continued and increased melt of mountain glaciers and ice caps, the Greenland Ice Sheet, sea ice, an earlier melt of river and lake ice, and a decrease in snow cover will all impact the Arctic radiation budget. Albedo will decrease and more solar radiation will be absorbed by the ground or open water, leading to additional warming. Furthermore, a decrease in albedo resulting from a decrease in snow cover over land may result in a further change in albedo through a change in the vegetation in response to surface warming (Chapin et al. 2005; Euskirchen et al. 2009). Permafrost thaw is also expected to lead to changes in albedo as areas dry out or become waterlogged. Indirectly, changes in surface albedo can modify large-scale atmospheric circulation (Dethloff et al. 2006).

For the Arctic glaciers (including those in Greenland surrounding the ice sheet), projected volume loss ranges between 12 and 32% of their current volume by 2100 depending on the GCM (Radić and Hock 2010). Any translation of ice volume losses into decreased areal extent will reduce albedo due to increases of bare ground. The thinning of the ice pack, earlier melt onset, and increased open water will contribute to a reduction in albedo. However, some recent research suggests that the ice-albedo feedback is potentially less efficient than previously thought (e.g., Graverson and Wang 2009). For example, the annual sea-ice minimum is reached in September at a time when incoming solar radiation is already weak.

The albedo of terrestrial snow-covered surfaces in the Arctic is a major feedback to climate over a large area (Fernandes et al. 2009), and an important feedback to climate globally through atmospheric and oceanic teleconnections. Recent observations show a general decrease in snow-cover extent and duration (AMAP 2011; Callaghan et al. 2011b [this issue]) that has led to decreases in the albedo feedback. Warming resulting from black carbon deposition on snow is equivalent to that resulting from a doubling of atmospheric CO_2 in Eurasia (Flanner et al. 2008; AMAP 2011).

The reflective properties of snow are strongly modified by vegetation (e.g., Essery et al. 2008; AMAP 2011; Callaghan et al. 2011b [this issue]). While new snow typically absorbs around 10% of thermal radiation incident on it, black spruce (*Picea mariana*) can absorb 95% of thermal radiation leading to a significant positive feedback (Juday et al. 2005). The snow feedback will be decreased as vegetation height increases above the snow pack during the ongoing and projected process of shrub and tree range extensions into the Arctic (Sturm et al. 2001; Tape et al. 2006). Although, increases in shrub advance in the Alaskan Arctic have not yet resulted in warming, it is predicted that an increase of shrubs could increase summer heating of the atmosphere by 3.7 W m^{-2}, which is equivalent to a doubling of CO_2 (Chapin et al. 2005). In the Barents region, changes in albedo through forest advance could increase temperatures by 1°C in spring (Gottel et al. 2008), although increased evapotranspiration and draw-down of atmospheric CO_2 in summer would result in cooling. Thus, feedbacks arising from changes in vegetation are complex, operating in different directions, through different mechanisms and over different periods of time.

Cloud Feedbacks

The cloud-radiation feedback in a warming climate is uncertain. Clouds both reflect solar radiation and absorb long-wave (terrestrial) radiation, the magnitudes of which depend on cloud amount, height, particle phase and size, and thickness. Some satellite studies (Wang and Key 2003; Liu et al. 2007) have shown that wintertime cloud amount in the Arctic appears to have been generally decreasing since the early 1980s, but springtime cloud amount has been increasing. While wintertime clouds in the Arctic have a warming effect, springtime cloud can have either a net warming or cooling effect. The overall radiative impact of these cloud cover changes on the surface is one of increased cooling. Therefore, changes in cloud cover may have actually suppressed Arctic warming to some degree (Wang and Key 2003).

The influence of changes in cloud cover on sea-ice extent and vice versa is an important part of the feedback process, but has not been studied yet extensively. On the time scale of a single season, changes in cloud amount may have minimal influence on summer sea-ice melt although there are clearly interdependencies between trends in cloud cover, surface temperature, and sea-ice extent. Over the past few decades, >80% of the observed surface warming in the western Arctic Ocean during autumn is attributable to decreasing sea ice. Similarly, over 80% of the winter surface cooling in the central Arctic is a result of changes in cloud cover. In spring, only about half of the surface warming is a result of changes in cloud cover (Liu et al. 2009).

Satellite and reanalysis data have shown that sea-ice retreat is linked to a decrease in low-level cloud amount and an increase in mid-level clouds (Schweiger et al. 2008). This is in contrast to the common notion that a warming ocean surface will increase surface evaporation and lead to more low clouds. While the response of cloud cover to sea-ice loss plays a minor role in regulating the summertime ice-albedo feedback, its role in the cloud-ice feedback in autumn is potentially larger (Kay and Gettelman 2009).

ATMOSPHERIC CIRCULATION

Overall, the atmosphere is a driver of change in the Arctic (Francis et al. 2009). It primarily forces rather than responds to changes in the cryosphere, a good example being changes in atmospheric wind patterns over the past decade that have contributed to recent reductions in summer Arctic sea-ice extent (Ogi et al. 2010). However, the increase in late summer open water area has, in turn, directly contributed to a modification of large-scale atmospheric circulation patterns. With a reduction in sea-ice cover in late summer, additional heat that was stored in the ocean is then released to the atmosphere in autumn. In years with reduced sea-ice cover the lower-tropospheric thickness is greater. This has a large-scale impact, even into the northern mid-latitudes, as the pressure fields and therefore winds are directly related to the atmospheric thickness. Responses can be complex, as the loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia.

There is also a relationship between snow cover and atmospheric circulation. Recent observational studies show that above-normal winter snow depth over western Russia and a corresponding below-normal snow depth over central Siberia—the east–west snow dipole—are associated with reduced Indian monsoon rainfall and above-normal sea-surface temperatures (SSTs) over the eastern and central tropical Pacific Ocean during subsequent winters (e.g., Peings and Douville 2010). Similarly, below-normal winter snow depth over European Russia and a corresponding above-normal snow depth over central Siberia are associated with increased monsoon rainfall and below-normal SSTs.

The large-scale effect of changes in snow cover through the snow albedo feedback (SAF) has recently been examined. For example, Fletcher et al. (2009) demonstrated a non-local influence of SAF on the summertime circulation in the extratropical Northern Hemisphere. In models with stronger SAF, increased land surface warming is associated with large-scale sea level pressure anomalies over the northern oceans and a poleward intensified subtropical jet. This would result in a change in heat and moisture fluxes into/out of the Arctic, although the feedback on snow cover and the impact on sea ice are not clear.

Ice sheets and ice caps change the elevation and form of the Earth's surface. This changes the air temperature and deforms the atmospheric circulation, forming a "wave shadow". For this reason, the Icelandic Low exists throughout the entire year, with the Greenland Ice Sheet to the west, as opposed to the Aleutian Low, which disappears in summer. Recent modeling supports the influence of the Greenland Ice Sheet on atmospheric circulation (Lunt et al. 2004). Due to the cooling effect of the Greenland Ice Sheet, the air temperature drops as much as 5–10°C in the atmospheric layer several 100 m above it. Over smaller ice caps and large mountain glaciers, the cooling is about 2–3°C and spreads upward to 200–250 m. This may affect the cyclone trajectories ("storm tracks") and the life cycle of pressure systems (Krenke 1982).

A retreating sea-ice margin may enhance melting over the Greenland Ice Sheet. Rennermalm et al. (2009) explored the spatial and temporal covariance of sea-ice extent and ice sheet surface-melt around Greenland from 1979 to 2007. Significant covariance was found in western Greenland. An examination of wind direction patterns and a lag analysis of ice retreat/advance and surface-melt event timings suggested that a change in sea-ice extent is a potential driver of ice-sheet melt, in that late summer wind directions bring onshore advection of ocean heat, enhanced by reductions in offshore sea ice. There is also a strong linkage between sea-ice loss and terrestrial permafrost temperature. Lawrence et al. (2008) found that the accelerated warming signal of rapid sea-ice loss penetrates up to 1500 km inland and substantially increases ground heat accumulation.

INTERACTIONS BETWEEN THE CRYOSPHERE AND THE FRESHWATER BUDGET OF THE ARCTIC

All of the cryospheric components play, to varying degrees, roles in the freshwater budget of the Arctic. Changes in the cryosphere can affect the strength of the Atlantic Meridional Overturning Circulation (AMOC), and hence, global climate. Assessing the magnitude of these effects requires not only an understanding of how the thermohaline circulation responds to freshwater inputs, but also the sources, locations, distributions, and pathways of freshwater into and out of the Arctic (Randall et al. 2007; On-line supplementary material).

In general, the Arctic Ocean receives freshwater inputs from direct precipitation, Pacific water through the Bering Strait (referenced to a particular salinity), terrestrial ice masses, and runoff from river basins. Importantly, the major river basins contributing flow to the Arctic Ocean are of a nival regime (runoff is dominated by snowmelt). Within the Arctic Ocean, freshwater amounts also change due to losses from evaporation and to the growth (−) and ablation (+) of sea ice. Very large volumes of freshwater can also be "stored" in deep basins with highly variable residence times. Freshwater export from the Arctic Ocean occurs primarily through Fram Strait and the Canadian Archipelago to the Atlantic Ocean, where it plays a role in the formation of deep water in the Greenland-Iceland-Norwegian (GIN) Seas.

It is the export of freshwater that has been identified within paleo-records as weakening the thermohaline circulation and causing major cooling events over the North Atlantic. For example, an Arctic Ocean pathway fed by freshwater flows from the Mackenzie River-Beaufort Gyre, instead of the previously supposed Great Lakes-St. Lawrence routing, has been recently identified as being responsible for the shutting down of the North Atlantic thermohaline circulation, resulting in major cooling associated with the Younger Dryas Period (e.g., Murton et al. 2010).

Since publication of the Arctic Climate Impact Assessment in 2005 (ACIA 2005), a number of updates have been made concerning the relative size of the overall freshwater budget terms and of the cryospheric components that contribute to them. These are summarized here and considered in the "On-line supplementary material", gleaned from the SWIPA Assessment (AMAP 2011) and related literature.

Recent Budget Estimates

Total freshwater inputs to the Arctic Ocean are dominated by river flow, inflow through the Bering Strait, and precipitation–evaporation directly occurring on the Arctic Ocean (Serreze et al. 2006; On-line supplementary material). Importantly, their estimate is an order of magnitude smaller than the total amount stored in the Arctic Ocean. Freshwater exports from the Arctic Ocean occur principally through the straits of the Canadian Arctic Archipelago and via Fram Strait as liquid and sea ice. Serreze et al. (2006) noted a larger freshwater inflow through Bering Strait and larger liquid freshwater outflow through Fram Strait than earlier estimates by others.

Peterson et al. (2006) conducted an analysis of changes in freshwater budget components for a broader "Arctic region" including the additional large land–ocean catchment of Hudson Bay in North America as well as the Nordic Seas and North Atlantic subpolar basins. Increasing precipitation–evaporation over the marine environments and larger river flow, probably also tied to increases in high-latitude precipitation, was estimated to have contributed $\sim$20 000 km^3 of freshwater to the total region from lows in the 1960s to highs in the 1990s (On-line supplementary material).

Freshwater Budget Components and Changes

Snowmelt and River Discharge

Compared to all other oceans, the Arctic Ocean receives a disproportionately large amount of river runoff to its total volume via the Lena, Mackenzie, Ob, and Yenisey rivers that are dominantly nival rivers. River flow provides the largest input to the Arctic Ocean freshwater budget (Prowse and Flegg 2000). Some increases in the magnitude and advances in timing of the snowmelt freshet on northern rivers have been observed and greater future changes are expected (AMAP 2011). Thawing permafrost also changes flow pathways and storage, which are important to how river water is distributed, directed and/or stored in the Arctic Ocean (e.g., Cooper et al. 2008; Jones et al. 2008).

Small Mountain Glaciers, Ice Caps, and the Greenland Ice Sheet

The overall freshwater contribution from these sources (Dyurgerov and Carter 2004; On-line supplementary material) in a broadly defined, pan-Arctic drainage basin, is far less than contributed by the corresponding nine major rivers. However, there is a greater "positive change signal" from the glaciers than river discharge (Dyurgerov and Carter 2004). Some freshwater budget analyses contain little discussion about the role of the Greenland Ice Sheet, despite its strategic placement as a freshwater source in the North Atlantic. In addition to freshwater volume, the location of the input may also be important (Randall et al. 2007) and meltwater runoff from the ice sheet is potentially a major source of freshening that has not yet been included in relevant models (Randall et al. 2007).

Sea Ice

The Arctic Ocean is a salt rather than temperature stratified ocean and hence, sea-ice growth/ablation and ocean dynamics can be greatly modified by changes in freshwater. Carmack (2000) estimated the volume of ice forming and melting each year (On-line supplementary material) but sea ice has undergone significant recent changes in areal coverage and thickness. For some key episodic losses, sea-ice-bottom melt has been linked to solar heating of the upper ocean (Perovich et al. 2008). Bottom melt can also result from the loss of thermal insulation from the warmer Atlantic water provided by the surface layers of freshwater and cold halocline. The stability of these upper layers, particularly with enhanced vertical mixing, has been identified as a "key wild card" regarding future sea-ice loss (Serreze et al. 2007). Sea ice is also exported through Fram Strait along with sea-ice meltwater, but export of sea-ice meltwater seems to be the least likely to influence thermohaline circulation (Jones et al. 2008).

Ocean Storage and Pathways

The fate of sea-ice meltwater and other forms of freshwater are not simply direct export because the Arctic Ocean also holds significant freshwater in storage with variable releases. While about one-quarter of the total is held on shelves, the majority is in the Eurasian and Canada basins, the latter being the largest single freshwater storehouse in the Arctic Ocean. Estimates of this storage (like the other freshwater budget terms) vary in the literature, and are primarily due to changes in import–export to Canada Basin and, in the accuracy and ability to measure its content. Despite variations in its estimated volume (Carmack 2000; Yamamoto-Kawai et al. 2008; On-line supplementary material) it is generally accepted that the largest source of the average annual freshwater input to this and other freshwater storehouses in the Arctic Ocean (Yamamoto-Kawai et al. 2009) is river runoff which is just slightly smaller than the amount removed by sea-ice formation. Average export of ice and liquid freshwater from Canada Basin contributes $\sim$40% of the freshwater flux to the North Atlantic.

The storage values vary with time due to atmospheric circulation, which can control pathways of freshwater to/from storage basins as well as the storage/release from within the storage basins [i.e., Ekman pumping, which under anticyclonic (cyclonic) circulation stores (releases) freshwater]. Most recently, measurements from Canada and Makarov basins indicate that there has been a freshwater storage increase in all the deep basins of the western Arctic (On-line supplementary material). By contrast, the Eurasian Basin in the eastern Arctic and closer to the main export to the North Atlantic has experienced a loss but there is a net gain (McPhee et al. 2009; On-line supplementary material). This net gain is a significant increase being approximately four times the volume associated with the Great Salinity Anomaly (a near-surface pool of fresher-than-usual water tracked in the subpolar gyre currents from around 1968 to 1982, which affected regional climate) and similar in magnitude to the total 1981–1995 sea-ice

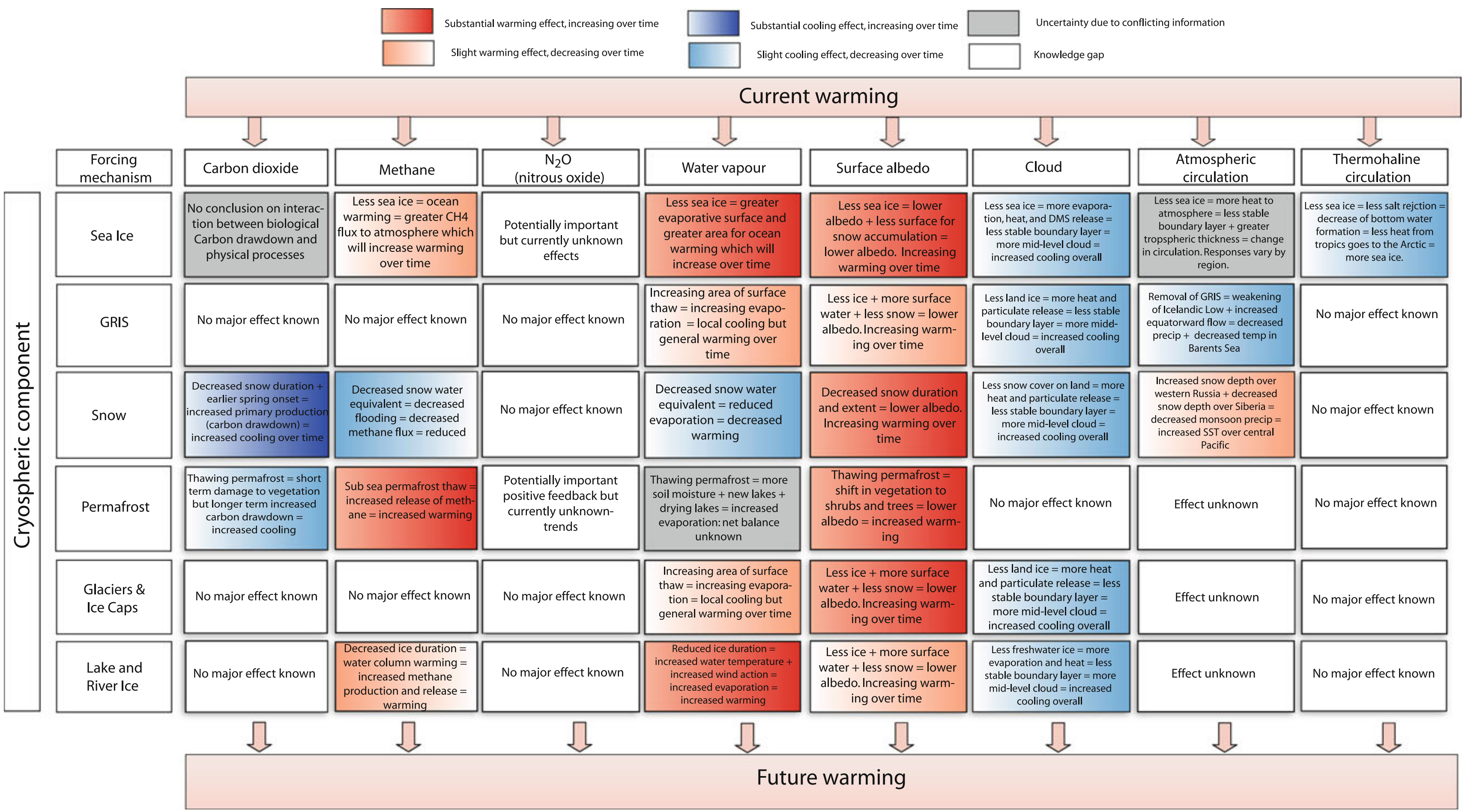

Fig. 2 The major feedbacks on the climate via the cryosphere. The *box colors* indicate the expected future impact of a change in the variable or process on the cryosphere (where *red* indicates warming and *blue* indicates cooling). The intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact or levels of understanding (where *gray* represents uncertainties due to conflicting information etc. and *white* indicates a knowledge gap). The change in intensity of colors from left to right indicates an acceleration or retardation of the process assuming a time horizon over ~30 years. The feedbacks presented in this figure represent a relatively simple synthesis of many complex processes

attrition (melt plus export) estimated in the above noted freshwater budget by Peterson et al. (2006).

Model Projections

At the time of the Arctic Climate Impact Assessment (ACIA 2005), there was a concern about the intensification of the hydrological cycle at high latitudes and the effect this would have on the AMOC. However, a strong scientific debate remains about the potential significance of freshwater effects on the AMOC (Randall et al. 2007). For example, Holland et al. (2007) noted that a constituent result among models for the period 1950–2050 (observations and modeled results) is an acceleration of the hydrological cycle, including increased ocean net-precipitation, river runoff, and net sea-ice melt. Together with Koenigk et al. (2007), they also noted, for liquid water, a larger export to lower latitudes, primarily through Fram Strait, and storage in the Arctic Ocean. By contrast, export and storage of freshwater in the form of sea ice decreases, although there is significant variability in sea-ice budget terms.

The Coupled Model Intercomparison Project (CMIP) more accurately defined the role of freshwater in AMOC weakening. All models used in CMIP show that AMOC weakening projected for the twenty-first century is caused more by changes in the surface heat flux than by freshwater, although its effect on high-latitude stratification plays a contributing role (Arzel et al. 2008). Also, interannual exchanges in freshwater between the GIN Seas and the North Atlantic were found to have a major driving influence on the interannual variability of deep convection over the twenty-first century.

SYNTHESIS

This review and analysis has substantiated and quantified many of the linkages between the cryosphere and climate identified by earlier recent studies (e.g., Francis et al. 2009). Figure 2 synthesizes the major feedbacks on climate mediated by cryospheric processes, where the impacts of changes in various forcing variables (e.g., albedo) and processes (e.g., atmospheric circulation) on elements of the cryosphere are depicted. The surface or near-surface air temperature is the variable primarily used to drive changes in the cryospheric component and then to assess the magnitude of warming or cooling. The box colors indicate the expected future impact of a change in the variable or process on the climate, where red indicates warming and blue indicates cooling. The colors also indicate the type of feedback, where red is a positive feedback from current climate warming to future temperature and blue is a negative feedback. The color does not, however, indicate the direction of change between the two variables alone. For example, a decrease in sea-ice cover increases atmospheric water vapor, which in turn may decrease sea-ice cover even further because water vapor is a greenhouse gas. This is a positive feedback, but the change in sea-ice cover and water vapor are in opposite directions. The intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact. For example, warming caused by a decrease in albedo through a loss of sea ice is expected to be greater than that due to changes in glacier extent. The feedbacks presented in Fig. 2 represent a relatively simple synthesis of many complex processes. This synthesis is intended to stimulate a better understanding of the processes rather than give a definitive analysis.

The impacts, interactions, and feedbacks in Fig. 2 are not necessarily independent. In particular, cloud formation is strongly dependent upon the available water vapor, which may be from local sources or transported from lower latitudes (atmospheric circulation). In addition, changes in one part of the cryosphere may cause changes in others.

CONCLUSIONS

Despite their potential importance, most of the terrestrial feedbacks are poorly quantified over large areas although considerable detail is available for a few specific locations. Generalization of the impacts is difficult at best. Large uncertainties exist in cloud feedbacks and subsea permafrost in particular; these topics require enhanced and sustained research. General circulation models currently include few feedbacks while the lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs.

The primary source of uncertainty regarding the atmosphere as a driver of change in the Arctic is precipitation. Future changes in its spatial and seasonal distribution are unclear, as are its effects on the sea-ice mass budget, marine primary productivity, and vegetation. A better understanding of precipitation effects will only be gained through developing and sustaining a more robust observing network.

There remains considerable controversy about the degree to which current levels of freshwater within the Arctic Ocean can affect the strength of the AMOC. There is therefore, a need to evaluate the potential for the cumulative production and release of large amounts of freshwater from all contributing components, including all components of the cryosphere, and particularly the ultimate freshwater export to the North Atlantic where it could produce a significant effect on the AMOC and global climate.

Acknowledgments We thank Sebastian Gerland, Mats Granskog, and two reviewers for comments on the manuscript. We also wish to thank the various national funding agencies that made this review possible, specifically the Swedish Environmental Protection Agency (grant number 08/210) and the Swedish Research Council FORMAS (grant number 204-2009-45) who helped directly support this publication together with Environment Canada. Prof. F. Wrona kindly facilitated and encouraged this publication. We also gratefully acknowledge the Arctic Monitoring and Assessment Program Secretariat and the SWIPA Integration Team in helping to produce the original SWIPA report and the various contributing SWIPA chapter authors.

REFERENCES

ACIA. 2005. *Arctic Climate Impact Assessment*, 1042 pp. Cambridge: Cambridge University Press.

Alekseev, G.V., and A.P. Nagurny. 2007. Role of sea ice in the formation of annual carbon dioxide cycle in the Arctic. *Doklady Earth Sciences* 417A: 1398–1401.

AMAP. 2011. *Snow, water, ice and permafrost in the Arctic (SWIPA)*. Arctic Monitoring and Assessment Programme (AMAP), Oslo.

Arzel, O., T. Fichefet, H. Goosse, and J.-L. Dufresne. 2008. Causes and impacts of changes in the Arctic freshwater budget during the twentieth and twenty-first centuries in an AOGCM. *Climate Dynamics* 30: 37–58.

Bolin, B. 2007. *A history of the science and politics of climate change. The role of the Intergovernmental Panel on Climate Change*, 292. Cambridge, UK: Cambridge University Press.

Callaghan, T.V., M. Johansson, J. Key, T. Prowse, M. Ananicheva, and A. Klepikov. 2011a. Feedbacks from changes in the Arctic cryosphere. In *Snow, water, ice and permafrost in the Arctic (SWIPA)*, Chap. 11.1. Arctic Monitoring and Assessment Programme (AMAP), Oslo.

Callaghan, T.V., M. Johansson, R.D. Brown, P. Ya. Groisman, N. Labba, V. Radionov, R.G. Barry, et al. 2011b. The changing face of Arctic snow cover: a synthesis of observed and projected changes. In *Arctic cryosphere—Changes and impacts*, ed. T.V. Callaghan, M. Johansson, and T.D. Prowse. *Ambio* 40(S1). doi: 10.1007/s13280-011-0212-y.

Carmack, E.C. 2000. The Arctic Ocean's freshwater budget: Sources, storage and export. In *The freshwater budget of the Arctic Ocean*, ed. Lewis E.L., E.P. Jones, P. Lemke, T.D. Prowse, and P. Wadhams, 91–126. Dordrecht: Kluwer Academic Publishers.

Chapin, F.S., M. Sturm, M.C. Serreze, J.P. McFadden, J.R. Key, A.H. Lloyd, A.D. McGuire, T.S. Rupp, et al. 2005. Role of land-surface changes in Arctic summer warming. *Science* 310: 657–660.

Christensen, T.R., T. Johansson, N. Malmer, J. Åkerman, T. Friborg, P. Crill, M. Mastepanov, and B. Svensson. 2004. Thawing subarctic permafrost: Effects on vegetation and methane emissions. *Geophysical Research Letters* 31: L04501. doi:10.1029/2003GL018680.

Cooper, L.W., J.W. McClelland, R.M. Holmes, P.A. Raymond, J.J. Gibson, C.K. Guay, and B.J. Peterson. 2008. Flow-weighted values of runoff tracers (d18O, DOC, Ba, alkalinity) from the six largest Arctic rivers. *Geophysical Research Letters* 35: L18606. doi:10.1029/2008GL035007.

Dethloff, K., A. Rinke, A. Benkel, M. Koltzow, E. Sokolova, S.K. Saha, D. Handorf, W. Dorn, et al. 2006. A dynamical link between the Arctic and the global climate system. *Geophysical Research Letters* 33: L03703.

Dorrepaal, E., S. Toet, R.S.P. van Logtestijn, E. Swart, M.J. van de Weg, T.V. Callaghan, and R. Aerts. 2009. Carbon respiration from subsurface peat accelerated by climate warming in the subarctic. *Nature* 460: 616–619.

Dyurgerov, M.B., and C.L. Carter. 2004. Observational evidence of increases in freshwater inflow to the Arctic Ocean. *Arctic, Antarctic, and Alpine Research* 36: 117–122.

Elberling, B., H.H. Christiansen, and B.U. Hansen. 2010. High nitrous oxide production from thawing permafrost. *Nature Geoscience* 3: 332–335.

Essery, R.L.H., J. Pomeroy, C. Ellis, and T. Link. 2008. Modelling longwave radiation to snow beneath forest canopies using hemispherical photography or linear regression. *Hydrological Processes* 22: 2788–2800.

Euskirchen, E.S., A.D. McGuire, D.W. Kicklighter, Q. Zhuang, J.S. Clein, R.J. Dargaville, D.G. Dye, J.S. Kimball, et al. 2006. Importance of recent shifts in soil thermal dynamics on growing season length, productivity, and carbon sequestration in terrestrial high-latitude ecosystems. *Global Change Biology* 12: 731–750.

Euskirchen, E.S., A.D. McGuire, F.S. Chapin III, S. Yi, and C.C. Thompson. 2009. Changes in vegetation in northern Alaska under scenarios of climate change, 2003–2100: Implications for climate feedbacks. *Ecological Applications* 19: 1022–1043.

Fernandes, R., H. Zhao, X. Wang, J. Key, X. Qu, and A. Hall. 2009. Controls on Northern Hemisphere snow albedo feedback quantified using satellite Earth observations. *Geophysical Research Letters* 36: 21. doi:10.1029/2009GL040057.

Flanner, M.G., C.S. Zender, P.G. Hess, N.M. Mahowald, T.H. Painter, V. Ramanathan, and P.J. Rasch. 2008. Springtime warming and reduced snow cover from carbonaceous particles. *Atmospheric Chemistry and Physics* 9: 2481–2497. doi:10.5194/acp-9-2481-2009.

Fletcher, C.G., P.J. Kushner, A. Hall, and X. Qu. 2009. Circulation responses to snow albedo feedback in climate change. *Geophysical Research Letters* 36: L09702.

Forster, P., V. Ramaswamy, P. Artaxo, T. Berntsen, R. Betts, D.W. Fahey, J. Haywood, J. Lean, et al. 2007. Changes in atmospheric constituents and in radiative forcing. In *Climate change 2007: The physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change*, ed. Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, and H.L. Miller. Cambridge, UK and New York, NY: Cambridge University Press.

Francis, J., D. White, J. Cassano, W. Gutowski Jr., L. Hinzman, M. Holland, M. Steele, and C. Vorosmarty. 2009. An arctic hydrologic system in transition: Feedbacks and impacts on terrestrial, marine, and human life. *Journal of Geophysical Research* 114: G04019. doi:10.1029/2008JG000902.

Gottel, H., J. Alexander, E. Keup-Thiel, D. Rechid, S. Hagemann, T. Blome, A. Wolf, and D. Jacob. 2008. Influence of changed vegetation fields on regional climate simulations in the Barents Sea Region. *Climatic Change* 87: 35–50.

Graverson, R.G., and M. Wang. 2009. Polar amplification in a coupled climate model with locked albedo. *Climate Dynamics* 33: 629–643.

Holland, M.M., J. Finnis, A.P. Barrett, and M.C. Serreze. 2007. Projected changes in Arctic Ocean freshwater budgets. *Journal of Geophysical Research* 112: G04S55. doi:10.1029/2006JG000354.

Jones, E.P., L.G. Anderson, S. Jutterström, L. Mintrop, and J.H. Swift. 2008. Pacific freshwater, river water and sea ice meltwater across Arctic Ocean basins: Results from the 2005 Beringia Expedition. *Journal of Geophysical Research* 113: C08012. doi: 10.1029/2007JC004124.

Jonsson, A., J. Karlsson, and M. Jansson. 2003. Sources of carbon dioxide supersaturation in clearwater and humic lakes in northern Sweden. *Ecosystems* 6: 224–235.

Juday, G.P, V. Barber, P. Duffy, H. Linderholm, S. Rupp, S. Sparrow, E. Vaganov, J. Yarie, et al. 2005. Forests, land management, and agriculture. ACIA. In *Arctic Climate Impacts Assessment:*

Technical report to the Arctic Council of Ministers, Chap. 14. Cambridge: Cambridge University Press.

Kay J.E., and A. Gettelman. 2009. Cloud influence on and response to seasonal Arctic sea ice loss. *Journal of Geophysical Research* 114: D18204. doi:10.1029/2009JD011773.

Koenigk, T., U. Mikolajewicz, H. Haak, and J. Jungclaus. 2007. Arctic freshwater export in the 20th and 21st centuries. *Journal of Geophysical Research* 112: GS04S41. doi:10.1029/2006JG000274.

Krenke, A.N. 1982. *Mass exchange in glacier systems over the USSR territory*. Leningrad: Hydrometeoizdat.

Lawrence, D.M., A.G. Slater, R.A. Tomas, M.M. Holland, and C. Deser. 2008. Accelerated Arctic land warming and permafrost degradation during rapid sea ice loss. *Geophysical Research Letters* 35: L11506. doi:10.1029/2008GL033985.

Liu, Y., J. Key, J. Francis, and X. Wang. 2007. Possible causes of decreasing cloud cover in the Arctic winter, 1982–2000. *Geophysical Research Letters* 34: L14705. doi:10.1029/2007GL030042.

Liu, Y., J. Key, and X. Wang. 2009. Influence of changes in sea ice concentration and cloud cover on recent Arctic surface temperature trends. *Geophysical Research Letters* 36: L20710. doi:10.1029/2009GL040708.

Lunt, D.J., N. de Noblet-Ducoudre, and S. Charbit. 2004. Effects of a melted Greenland ice sheet on climate, vegetation, and the cryosphere. *Climate Dynamics* 23: 679–694.

Mack, M.C., M.S. Bret-harte, T.N. Hollingworth, R.R. Jandt, E.A.G. Schuur, G.R. Shaver, and D.L. Verbyla. 2011. Carbon loss from an unprecedented Arctic tundra wildfire. *Nature* 475: 489–492.

Mastepanov, M., C. Sigsgaard, E.J. Dlugokencky, S. Houweling, L. Ström, M.P. Tamstorf, and T.R. Christensen. 2008. Large tundra methane burst during onset of freezing. *Nature* 456: 628–631.

McGuire, A.D., L.G. Anderson, T.R. Christensen, S. Dallimore, L. Guo, D.J. Hayes, M. Heimann, T.D. Lorenson, et al. 2009. Sensitivity of the carbon cycle in the Arctic to climate change. *Ecological Monographs* 79: 523–555.

McPhee, M.G., A. Proshutinsky, J.H. Morison, M. Steele, and M.B. Alkire. 2009. Rapid change in freshwater content of the Arctic Ocean. *Geophysical Research Letters* 36: L10602. doi:10.1029/2009GL037525.

Murton, J.B., M.D. Bateman, S.R. Dallimore, J.T. Teller, and Z. Yang. 2010. Identification of Younger Dryas outburst flood path from Lake Agassiz to the Arctic Ocean. *Nature* 464: 740–743. doi:10.1038/nature08954.

Ogi, M., K. Yamazaki, and J.M. Wallace. 2010. Influence of winter and summer surface wind anomalies on summer Arctic sea ice extent. *Geophysical Research Letters* 37: L07701. doi:10.1029/2009GL042356.

Peings, Y., and H. Douville. 2010. Influence of the Eurasian snow cover on the Indian summer monsoon variability in observed climatologies and CMIP3 simulations. *Climate Dynamics* 34: 643–660.

Perovich, D.K., J.A. Richter-Menge, K.F. Jones, and B. Light. 2008. Sunlight, water, ice: Extreme Arctic sea melt during the summer of 2007. *Geophysical Research Letters* 35: L11501. doi:10.1029/2008GL034007.

Peterson, B.J., J. McClelland, R. Curry, R.M. Holmes, J.E. Walsh, and K. Aagaard. 2006. Trajectory shifts in the Arctic and Subarctic freshwater cycle. *Science* 313: 1061–1066.

Prowse, T.D. 2009. Introduction: hydrologic effects of a shrinking cryosphere. *Hydrological Processes* 23: 1–6. doi:10.1002/hyp.7215.

Prowse, T.D., and P.O. Flegg. 2000. The magnitude of river flow to the Arctic Ocean: Dependence on contributing area. *Hydrological Processes* 14: 3185–3188.

Radić, V., and R. Hock. 2010. A new estimate of regional and global volumes of glaciers derived from statistical up-scaling of glacier inventory data. *Journal of Geophysical Research* 115: F01010. doi:10.1029/2009JF001373.

Randall, D.A., R.A. Wood, S. Bony, R. Colman, T. Fichefet, J. Fyfe, V. Kattsov, A. Pitman, et al. 2007. Climate models and their evaluation. In *Climate change 2007: The physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change*, ed. Solomon, S., D. Quin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, and H.L. Miller. Cambridge, UK and New York, NY: Cambridge University Press.

Rennermalm, A.K., L.C. Smith, J. Stroeve, and V. Chu 2009. Does sea ice influence Greenland ice sheet surface-melt? *Environmental Research Letters* 4: 024011. doi:10.1088/1748-9326/4/2/024011.

Repo, M.E., S. Susiluoto, S.E. Lind, S. Jokinen, V. Elsakov, C. Biasi, T. Virtanen, and P.J. Martikainen. 2009. Large N_2O emissions found from cryoturbated peat soil in Tundra. *Nature Geoscience* 2: 189–192.

Rouse, W.R., M.S.V. Douglas, R.E. Hecky, A.E. Hershey, G.W. Kling, L. Lesack, P. Marsh, M. McDonald, et al. 1997. Effects of climate change on the freshwaters of Arctic and subarctic North America. *Hydrological Processes* 11: 873–902.

Schuur, E.A.G., J.G. Vogel, K.G. Crummer, H. Lee, J.O. Sickman, and T.E. Osterkamp. 2009. The impact of permafrost thaw on old carbon release and net carbon exchange from tundra. *Nature* 459: 556–559.

Schweiger, A.J., R.W. Lindsay, S. Vavrus, and J.A. Francis. 2008. Relationships between Arctic sea ice and clouds during autumn. *Journal of Climate* 21: 4799–4810.

Scinocca, J.F., M.C. Reader, D.A. Plummer, M. Sigmond, P.J. Kushner, T.G. Shepherd, and A.R. Ravishankara. 2009. Impact of sudden Arctic sea-ice loss on stratospheric polar ozone recovery. *Geophysical Research Letters*. doi:10.1029/2009GL041239.

Serreze, M.K., A.P. Barrett, A.G. Slater, R.A. Woodgate, K. Aagaard, R.B. Lammers, M. Steele, R. Moritz, M. Meredith, and C.M. Lee. 2006. The large-scale freshwater cycle of the Arctic. *Journal of Geophysical Research* 111: C11010. doi:10.1029/2005JC003424.

Serreze, M.C., M.M. Holland, and J. Stroeve. 2007. Perspectives on the Arctic's shrinking sea-ice cover. *Science* 315: 1533–1536.

Shakhova, N., I. Semiletov, A. Salyuk, V. Yusupov, D. Kosmach, and O. Gustafsson. 2010. Extensive methane venting to the atmosphere from sediments of the East Siberian Arctic Shelf. *Science* 327: 1246–1250.

Shindell, D., G. Faluvegi, A. Lacis, J. Hansen, R. Ruedy, and E. Aguilar. 2006. Role of tropospheric ozone increases in 20th century climate change. *Journal of Geophysical Research* 111: D08302. doi:10.1029/2005JD006348.

Smith, L.C., Y. Sheng, G.M. MacDonald, and L.D. Hinzman. 2005. Disappearing Arctic lakes. *Science* 308: 1429.

Sturm, M., C. Racine, and K. Tape. 2001. Climate change: Increasing shrub abundance in the Arctic. *Nature* 411: 546–547.

Tape, K., M. Sturm, and C. Racine. 2006. The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. *Global Change Biology* 12: 686–702.

Tarnocai, C., J. Canadell, E.A.G. Schuur, P. Kuhry, G. Mazhitova, and S. Zimov. 2009. Soil organic carbon stocks in the northern circumpolar permafrost region. *Global Biogeochemical Cycles* 23: GB2023. doi:10.1029/2008GB003327.

Voulgarakis, A., X. Yang, and J.A. Pyle. 2009. How different would tropospheric oxidation be over an ice-free Arctic? *Geophysical Research Letters* 36: L23807. doi:10.1029/2009GL040541.

Walter, K.M., L.C. Smith, and F.S. Chapin III. 2007. Methane bubbling from northern lakes: present and future contributions to the methane budget. *Philosophical Transactions of the Royal Society A* 365: 1657–1676. doi:10.1098/rsta.2007.2036.

Wang, X., and J. Key. 2003. Recent trends in Arctic surface, cloud, and radiation properties from space. *Science* 299: 1725–1728.

Yamamoto-Kawai, M., F.A. McLaughlin, E.C. Carmack, S. Nishino, and K. Shimada. 2008. Freshwater budget of the Canada Basin, Arctic Ocean, from salinity, $\delta^{18}0$, and nutrients. *Journal of Geophysical Research* 113: C01007. doi:10.1029/2006JC003858.

Yamamoto-Kawai, M., F.A. McLaughlin, E.C. Carmack, S. Nishino, K. Shimada, and N. Kurita. 2009. Surface freshening of the Canada Basin, 2003–2007: River runoff versus seas ice meltwater. *Journal of Geophysical Research* 114: C00A05. doi: 10.1029/2008JC005000.

AUTHOR BIOGRAPHIES

Terry V. Callaghan (✉) is a Distinguished Research Professor at the Royal Swedish Academy of Sciences and Professor of Arctic Ecology at the Universities of Sheffield, UK and Lund, Sweden. He specializes in arctic ecology, and climate and UV-B radiation impacts on arctic ecosystems.
Address: Royal Swedish Academy of Sciences, 104 05 Stockholm, Sweden.
e-mail: terry_callaghan@btinternet.com

Margareta Johansson is a Research Scientist at the Department of Earth and Ecosystem Sciences, Lund University and at the Royal Swedish Academy of Sciences, Stockholm, Sweden. She specializes in permafrost dynamics in relation to climate change and its impact on ecosystems.
Address: Division of Physical Geography and Ecosystem Analyses, Department of Earth and Ecosystem Sciences, Lund University, Sölvegatan 12, 223 62 Lund, Sweden.
e-mail: margareta.johansson@nateko.lu.se

Jeff Key is a Research Scientist with the US National Oceanic and Atmospheric Administration's Satellite and Information Service, and is Chief of the Advanced Satellite Products Branch. He specializes in satellite remote sensing of snow, ice, and clouds in the Polar Regions.
Address: NOAA/NESDIS, 1225 West Dayton Street, Madison, WI 53706, USA.
e-mail: Jeff.Key@noaa.gov

Terry Prowse is a senior Research Scientist with Environment Canada and a Professor in the Department of Geography at the University of Victoria. His research focuses on the effects of climate variability and change on cold-regions hydrology, hydro-climatology, and hydro-ecology.
Address: Environment Canada, Department of Geography, University of Victoria, Victoria, BC V8P 5C2, Canada.
e-mail: terry.prowse@ec.gc.ca

Maria Ananicheva is a senior Research Scientist in the Institute of Geography, Glaciological Department, Russian Academy of Sciences. Her area of interest is glacio-climatology, mountain glacier change study in the Russian Arctic, contemporary mass balance of glaciers and projection for the near future by AOGCM scenarios.
Address: Institute of Geography, Russian Academy of Sciences, Staromonetny per 29, Moscow, Russia 119017.
e-mail: maria_anan@rambler.ru

Alexander Klepikov is a Physical Oceanographer who works as a senior scientist at the Arctic and Antarctic Research Institute in St. Petersburg, Russia. He is currently the head of Department for the Antarctic Oceanography and Climate Studies. His research interests include physical oceanography, modeling of the Arctic and Southern Oceans, and polar climate dynamics.
Address: Arctic and Antarctic Research Institute, 38 Bering Street, St. Petersburg, Russia 199397.
e-mail: klep@aari.ru