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Trajectory of the Arctic as an integrated system

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Abstract. Although much remains to be learned about the Arctic and its component processes, many of the most urgent scientific, engineering, and social questions can only be approached through a broader system perspective. Here, we address interactions between components of the Arctic system and assess feedbacks and the extent to which feedbacks (1) are now underway in the Arctic and (2) will shape the future trajectory of the Arctic system. We examine interdependent connections among atmospheric processes, oceanic processes, sea-ice dynamics, marine and terrestrial ecosystems, land surface stocks of carbon and water, glaciers and ice caps, and the Greenland ice sheet. Our emphasis on the interactions between components, both historical and anticipated, is targeted on the feedbacks, pathways, and processes that link these different components of the Arctic system. We present evidence that the physical components of the Arctic climate system are currently in extreme states, and that there is no indication that the system will deviate from this anomalous trajectory in the foreseeable future. The feedback for which the evidence of ongoing changes is most compelling is the surface albedo-temperature feedback, which is amplifying temperature changes over land (primarily in spring) and ocean (primarily in autumn–winter). Other feedbacks likely to emerge are those in which key processes include surface fluxes of trace gases, changes in the distribution of vegetation, changes in surface soil moisture, changes in atmospheric water vapor arising from higher temperatures and greater areas of open ocean, impacts of Arctic freshwater fluxes on the meridional overturning circulation of the ocean, and changes in Arctic clouds resulting from changes in water vapor content.

Key words: Arctic atmosphere dynamics; Arctic climate change trajectories; Arctic climate system feedbacks; Arctic hydrology; Arctic land ecosystems; Arctic marine ecosystems; Arctic ocean dynamics; Arctic sea ice; Greenland ice sheet; permafrost.

Introduction

The Arctic has experienced unprecedented changes in recent decades. Table 1 presents a summary of the changes observed in individual variables discussed in this paper and an assessment of our ability to capture those changes in models. Permafrost is warming, glaciers are melting and retreating, hydrological processes are being altered; and biological and social systems are changing in response (e.g., Hinzman et al. 2005, Grebmeier et al. 2010, Tedesco et al. 2011, Mernild et al. 2012). Knowing how the structure and function of Arctic ecosystems respond to recent and persistent climate change is paramount to understanding the future state of the Earth system and how humans must adapt (U.S. Arctic Research Commission 2010). New extremes in seasonal surface climate conditions are being experienced. A range of biophysical states and processes, influenced by threshold and phase changes at the freezing point, are being altered. Hydrological and biogeochemical cycles are shifting. And, more frequently, human subsystems are being affected. As in all ecosystems, the components of the Arctic system are interdependently linked to all others (Fig. 1) and inherently interconnected with the broader global system (Roberts et al. 2010, Saito et al. 2013). However, unlike most other ecosystems, winter conditions dominate in the Arctic for most or all of the year. When thawing of the snow, ice, and near-surface soil or water occurs, dramatic changes manifest through threshold shifts in surface energy balance, glaciological and hydrological processes, and ecosystem dynamics. Such significant changes reverberate throughout the system, particularly if these areas do not usually experience phase change or if the thawed period is lengthened. Also playing roles in the broad-scale changes

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TABLE 1. Summary of environmentally and climatically important variables and observed trends in recent decades.

Parameter	Trajectory sign	Confidence	Degree of model agreement	Comments
Atlantic Water temperature	increase	high	not well known	Pan-Arctic impact on ice to be quantified.
Pacific Water temperature	increase	medium high	not well known	Pan-Arctic impact on ice to be quantified.
Arctic Ocean freshwater balance	increase	high	medium	Export of ice volume through straits needs refining.
Sea ice area/extent	decrease in summer/ potential increase in winter	high/medium low	high	Physical mechanisms of summer ice area increase need further clarification.
Sea ice thickness (or age)	decrease	high	high	Percent decrease greater than for area, extent.
Surface air temperature	increase	high to medium high	high	Relative role of local vs. global effects needs clarification.
Air humidity	increase	high to medium high	medium	Relative role of local vs. global drivers needs clarification.
Cloudiness	increase	medium	low	Increase in only some seasons.
Precipitation	increasing	high	high	Spatial distribution needs refining.
Windiness	uncertain	low	uncertain	No comprehensive evaluations available.
Global glacier and ice cap (GIC) mass balance	decrease	high	low	Few surface mass balance observations exist.
Greenland ice sheet melt extent	increase	high	medium	Spatial variability needs clarification.
Global sea-level rise due to GIC wastage	increase	high	medium	Raising mean global sea level by approximately 1.0 mm sea level extent/ yr.
Ice flow velocities and calving	increase	medium	low	Mechanisms of ice dynamics are poorly understood.
Permafrost temperature and thawing	increase	high	high	High uncertainty on future rate of permafrost degradation.
Active layer thickness (ALT)	variable	low	low	Conflicting processes influence ALT.
Coastal erosion rates	increase	high	low	Complex interaction of multiple processes.
Snowpack thickness	increase	medium	low	High spatial and temporal variability.
Evapotranspiration	increase	high	low	High spatial and temporal variability.
Lake or pond size and number	increase and decrease	high	low	Function of permafrost extent and water balance.
Wetland extent	decrease	high	low	Function of permafrost extent and water balance.
River discharge	increase	high	high	Spatial distribution needs refining.
Soil moisture	primarily decreasing, limited areas of increasing	high	variable	No consistent observations for significant change, but predictions are for widespread drying, but increased wetness in areas of groundwater upwelling.
Groundwater storage	variable responses	high	none	Measurements completed with GRACE demonstrate Lena and Yenisei watersheds, decreased in the Mackenzie watershed.
Terrestrial surface albedo	decrease	high	high	Greatest contribution from change in snow duration.

Table 1. Continued.

Parameter	Trajectory sign	Confidence	Degree of model agreement	Comments
Fire frequency	increase	high	high	Shift in boreal forest composition to deciduous forest may limit increase.
Change in surface energy balance	summer heating of the surface atmosphere	high	high	Greatest contribution from changes in snow duration.
Terrestrial–atmosphere exchange of carbon dioxide	shift from a terrestrial sink to a terrestrial source of carbon dioxide to the atmosphere	medium low	medium low	Models that explicitly link permafrost thaw to decomposition of permafrost soil carbon generally agree that terrestrial ecosystems will be a source of carbon dioxide to the atmosphere.
Terrestrial methane emissions	no current evidence of increasing emissions, but generally expectations for increase in the future	medium high	medium high	Future projections from models that methane emissions will increase are primarily driven by changes in temperature.
Carbon export from terrestrial ecosystems to the coastal ocean	no current evidence that carbon export is increasing, and there are competing theoretical models as to whether it will increase in the future	low	low	One school of thought is that permafrost thaw will lead to more water flux through mineral soil to decrease export. The other school of thought is that permafrost thaw will expose more carbon to the possibility of export.
Marine primary production (PP)	current evidence suggests PP may be decreasing in some areas; remote sensing and numerical models suggest recent increase overall	low	low	Three global climate carbon-cycle models project an increase and one a decrease in Arctic PP.
Distribution shifts of marine species	numerous examples recorded; many display trend northward	high to low	more model results needed for comparisons	Some models shows shifts for species at low trophic levels.
Ocean acidification and reduced saturation states of calcium carbonate minerals	increase	calcium carbonate mineral saturation state suppression observed in localized areas	medium high	Projections from two global coupled carbon-cycle climate models available.
Marine–Atmosphere exchange of carbon dioxide	increase sink of carbon dioxide to the atmosphere expected from enhanced primary production (new) and carbon export	medium	low	Near-term increasing. Long-term may be decreasing. Impact from other ocean and sea ice biological processes uncertain.
Marine source of biogenic aerosols and aerosol precursors that influence CCN properties	no current evidence of increase, but enhanced production with sea ice loss expected	low	not well known	Modeling of DMS more advanced than for other organics.

Note: Abbreviations are: CCN, cloud condensation nuclei; DMS, dimethylsulfide.

across the entire system are the closely coupled feedbacks that connect various system processes. The dynamic interactions among feedbacks affect numerous processes, each often considered the domain of a distinct academic discipline. To improve our understanding of the role of the Arctic in the global climate system, we must develop a more integrated view of the terrestrial, aquatic, oceanic, and atmospheric processes and their linkages. The papers in this Invited Feature focus upon the complex interplay of the components and processes that comprise the Arctic as a system. The component interactions addressed in this synthesis paper lead to important feedback processes involving the ocean, glacier ice, sea ice, marine and terrestrial ecosystems, and hydrologic and geophysical components of the Arctic system. These interactions must be quantified if we are to understand and predict the complex feedbacks that shape Arctic ecosystems.

The following sections contain a synthesis of work that in recent years has provided substantial insight into the dynamics of the Arctic system. While an assessment of feedbacks is an ultimate goal, our approach in this synthesis is to focus first on the processes and pathways of Arctic system components: the atmosphere and ocean, ice sheets and glaciers, and hydrology and permafrost. We then integrate these component processes and pathways into a broader discussion of feedbacks in Arctic terrestrial ecosystems and marine ecosystems, with an eye toward the global implications of Arctic ecosystem feedbacks. In essence, here we consider the key component processes and pathways as the building blocks of feedback processes that can amplify or attenuate Arctic change.

Key Atmosphere and Ocean Pathways

Atmosphere: observed changes and key pathways

Over the last decade, Arctic (>60° N) surface air temperatures (SATs) have been exceptionally high, with anomalies relative to 1961–2000 means reaching 1.72°C in 2003, 2.11°C in 2005, and 2.18°C in 2007, the warmest years in the history of instrumental observations (Bekryaev et al. 2010). These years were also associated with minimum Arctic ice cover. Further, recent Arctic warming has become more seasonally uniform compared to warming in the 1930s–1940s (Fig. 2). There has also been substantial spatiotemporal heterogeneity of the warming pattern, which may be associated with enhanced high-latitude warming, known as polar amplification (e.g., Serreze and Francis 2006, Serreze et al. 2009, Bekryaev et al. 2010). Accelerated Arctic warming in recent decades has been captured by recent SAT trend maps (Fig. 3; Bekryaev et al. 2010). For example, the Arctic warming trend over the last decade was exceptionally strong, reaching 1.35°C per decade. Arctic warming trends over longer periods of time were also very strong: up to 1.36°C per century for 1875–2008, almost twice as strong as Northern Hemisphere SAT trends (0.79°C per century). The strongest Arctic warming of the most recent decade occurred in autumn and early winter (Fig. 4) in response to sea ice retreat; a weaker maximum occurred in spring. The earlier loss of snow on land appears to have contributed to the springtime maximum.

Because polar amplification has enhanced the Arctic warming of recent decades, a discussion of key interactions involving atmospheric change must include polar amplification. The nature of polar amplification has been a focus of work over the past several years. Bekryaev et al. (2010) developed a new statistical measure of polar amplification, linking high-latitude and hemispheric SAT anomalies through a regression relationship and yielding a stable metric of polar amplification. This measure indicated that polar regions warmed at 1.6 times the rate of lower latitudes during the period of 1875–2008. Bekryaev et al. (2010) also show that polar amplification has two scales: hemispheric and local. The local contribution, related to the ice–albedo feedback mechanism, has been strongest in the autumn and in the marginal ice zone, which includes the northern coastal regions. However, heat budget estimates indicate that the recent high Arctic SATs cannot be explained by the ice–albedo feedback mechanism alone, and that the large-scale contribution cannot be ignored.

Evaluations of the recent Arctic temperature trends provide further evidence that advective transports are contributing to enhanced Arctic warming (Graversen et al. 2008, Serreze et al. 2009, Porter et al. 2012). For example, Graversen (2006) examined the linkage between atmospheric northward energy transport across 60°N and Arctic surface air temperatures. During 1979–2001, the atmospheric northward energy transport experienced an overall positive but weak trend and was largest during the period from the mid-1980s to the mid-1990s. Thus the various studies summarized here all point to a consistent interpretation, that polar amplification consists of two components: an advective contribution originating in transports from lower latitudes, and a local contribution in which the retreat of sea ice and snow triggers a feedback with near-surface air temperatures. The net result is a seasonal variation of the polar amplification (Fig. 4).

To a considerable extent, the recent spatial patterns of the temperature change have been shaped by the phase of low-frequency (decadal or multidecadal) variations of the atmospheric circulation. Two large-scale modes, of which there are documented effects on regional Arctic SATs, are (1) the Arctic Oscillation, which drives temperature anomalies from eastern Canada across the North Atlantic to northern Eurasia (Thompson and Wallace 2000), and (2) the Pacific Decadal Oscillation, which is a pattern of Pacific climate variability and has a strong influence on Subarctic temperatures in the Pacific sector (Mantua and Hare 2002). The Arctic warming of the late 1980s and early 1990s has been attributed to a predominantly positive phase of the Arctic Oscillation

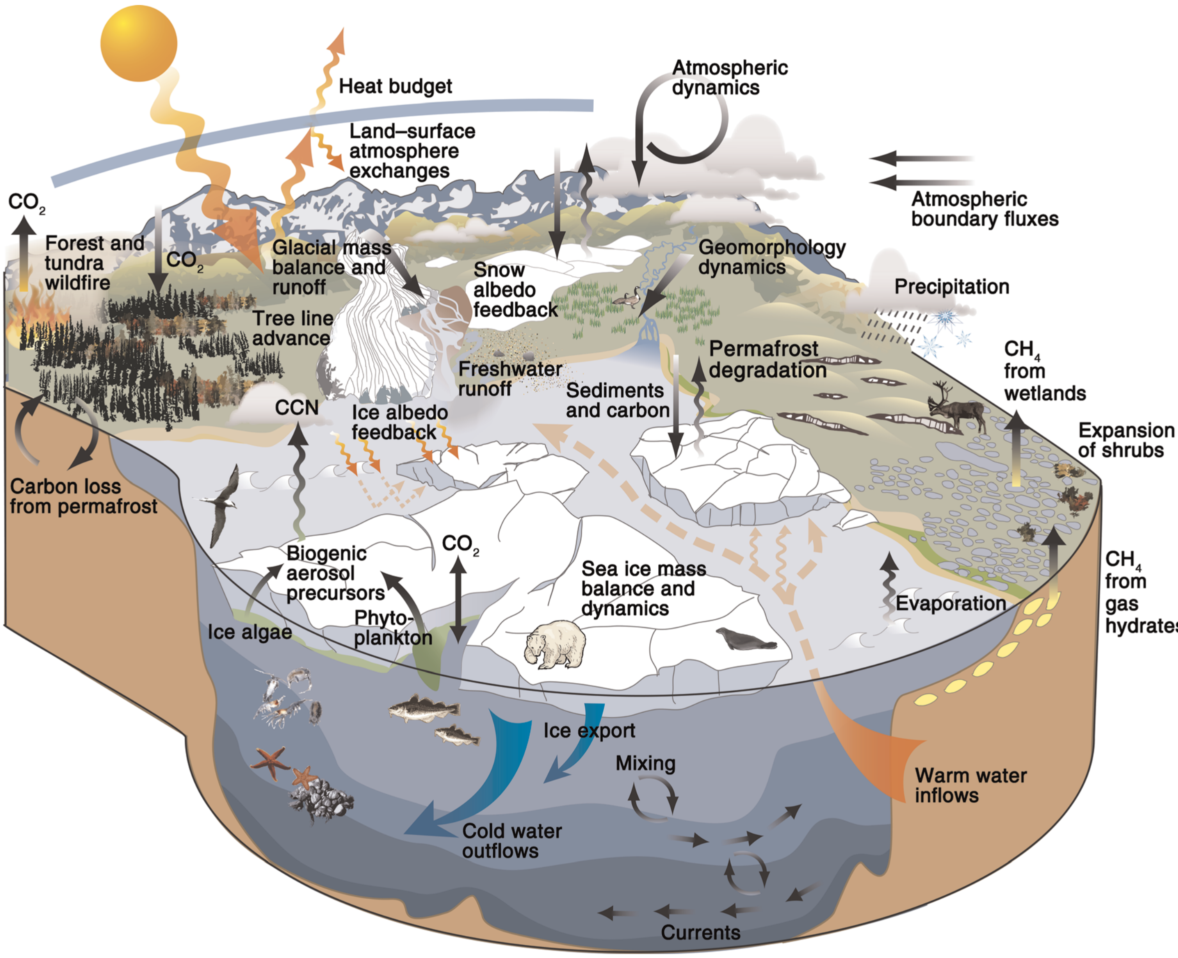

FIG. 1. The Arctic behaves as a tightly coupled system, strongly influenced by feedback pathways among the system components. These linkages among system components complicate our ability to adequately characterize the environmental responses to a warming climate. CCN stands for cloud condensation nuclei.

(e.g., Comiso 2003, Overland et al. 2008). Indeed, the warming of the 1980–1990s was stronger over northern Eurasia than over many other sectors of the Arctic. In contrast, more recent Arctic warming cannot be attributed to the Arctic Oscillation. First, the recent Arctic Oscillation has been in a generally neutral state, oscillating between positive and negative phases since 1997; yet, the Arctic's warmest years in the instrumental record have occurred in the past decade (Fig. 2). Second, the Arctic Oscillation Index reached the most negative values ever recorded in two recent winters, 2009–2010 and 2010–2011. At the same time, the high Arctic was relatively warm.

Several recent studies have pointed to the prominent role of similar circulation patterns, assigned names ranging from "Dipole Anomaly" (Watanabe et al. 2006, Wang et al. 2009) to "Arctic Dipole" and "Arctic Rapid-Change Pattern" (Zhang et al. 2008). These patterns arguably preconditioned the Arctic sea-ice cover for the rapid summer retreat of the late 2000s (Smedsrud et al. 2008). Overland et al. (2008) highlight the meridional (across-pole) character of this atmospheric pattern, which has affected sea-ice export in a pattern distinct from the Pacific Decadal Oscillation and the Arctic Oscillation. In turn, there are indications that the loss of sea ice has impacted the atmospheric circulation during autumn and early winter (see *Sea ice: observed changes and key pathways*).

Ocean: observed changes and key pathways

Over recent decades, the state of the upper (<50–75 m) Arctic Ocean has experienced substantial changes, including the warming, thinning, or loss of sea ice and consequent effects to the biological system. Enhanced upper-ocean solar heating through leads and consequent ice-bottom melting may have contributed to these changes (Perovich et al. 2008, Toole et al. 2010). For example, summer uptake of solar heat by the upper ocean through diminishing ice area (as a manifestation of ice–albedo feedback) was estimated in the Beaufort Sea to be on the order of 100 W/m^2 (Perovich et al. 2008), further reducing Arctic ice cover. Observations

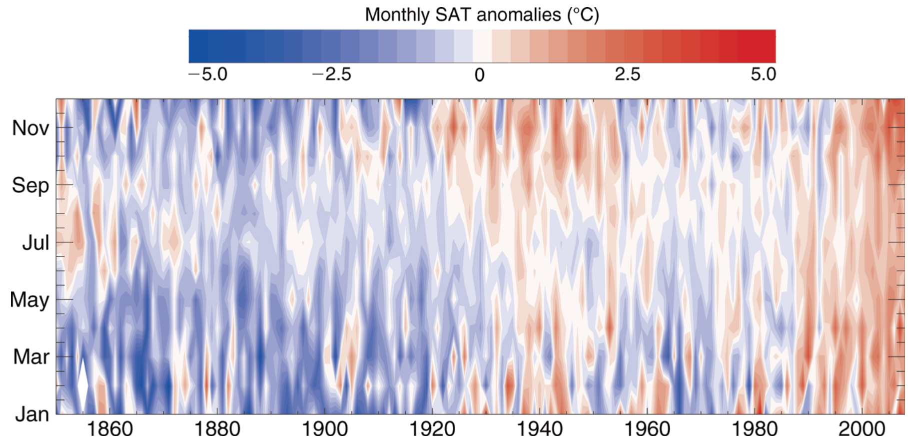

FIG. 2. Year vs. month Arctic (>60° N) SAT anomalies, relative to 1961–2000, derived from meteorological land stations. Over the last decade, Arctic warming has become stronger and more seasonally uniform compared to warming in the 1930s–1940s. From Bekryaev et al. (2010).

conducted under the Nansen and Amundsen Basins Observational System program carried out in September 2007 showed a substantial (up to 2°C or more) water temperature increase in the upper 25-m layer in the Siberian sector of the Arctic Ocean. Bekryaev et al. (2010) compared this anomalous upper-ocean heat uptake to the advective annual horizontal atmospheric heat transport along 60°N. They found that the standard deviation of atmospheric horizontal advection of heat was almost three times more than the total amount of heat accumulated in the ice-free area in summer 2007. Bekryaev et al. concluded that the dramatic reduction of Arctic ice in 2007, as well as anomalously high Arctic SATs, cannot be explained using the ice-albedo feedback mechanism alone, pointing to the importance of heat transport into the Arctic by atmospheric circulation and ocean currents. This conclusion agrees well with Graversen and Wang's (2009) modeling results, which show that the ice–albedo feedback mechanism is responsible for only 15% of polar amplification. Schweiger et al. (2008) used a regional coupled ice–ocean model to argue that an increase of short-wave radiation in response to a lack of clouds did not contribute significantly to the summer 2007 ice minimum; instead, they proposed that downward long-wave radiation accounted for a substantial fraction of Arctic ice loss.

The sea-ice reduction in the Canadian Basin of the Arctic Ocean that began in the late 1990s as a result of increased influx of warm summer waters of Pacific origin has clearly shown the thermodynamic coupling between Arctic ice and the ocean interior (i.e., below the upper mixed layer). Shimada et al. (2006) proposed that Pacific water, which enters the Arctic Ocean through the Bering Strait and occupies the 20–100 m depth range, has played a fundamental role in the recent sea-ice reduction in the Canada Basin of the Arctic Ocean. Warming of the Pacific water layer has retarded sea-ice formation during winter, and an imbalance between formation and melting has caused sustained loss of sea ice in this area since the late 1990s. The same conclusion was made by Woodgate et al. (2010), who analyzed Pacific water transports through the Bering Strait. Woodgate et al. found a doubling of heat flux from 2001 through 2007: enough to explain a third of the 2007 summer Arctic ice thickness loss.

Another gateway supplies the Arctic Ocean with Atlantic water through the Fram Strait. Over the 20th century, a temperature record of the Arctic Ocean intermediate (depth range ~150–900 m), warm (temperature >0°C), and salty water of Atlantic origin (usually called Atlantic Water, AW) is characterized by two warm periods (in the 1930s–1940s and in recent decades), and two cold periods (early in the 20th century and in the 1960s–1970s; Fig. 5; Polyakov et al. 2004). AW warming over the recent past has not been steady over time, as approximately decadal-scale pulses of warm AW have penetrated into the Arctic Ocean interior (e.g., Schauer et al. 2004, Polyakov et al. 2005). The latest warm pulse of exceptional strength, which entered the Arctic Ocean in the mid-2000s, moved through the Eurasian Basin and downstream into the Arctic Ocean interior (e.g., Schauer et al. 2004, 2011, Dmitrenko et al. 2008; see also Figs. 2 and 7 and discussion in Polyakov et al. 2013). However, very recent oceanographic surveys and mooring observations demonstrate that the recent 2006–2007 temperature peak was followed by cooling. This cooling likely contributed to the 2008–2009 reduction of the warm AW pulse from the eastern Eurasian Basin.

Analysis of freshwater content (FWC) anomalies observed in 2007 (a year with exceptional data coverage, thanks to the International Polar Year) demonstrates that the western Arctic Ocean was saltier than normal, whereas the eastern Arctic Ocean was fresher. Analyzing long-term, high-latitude FWC changes, Polyakov et al. (2008) suggested that ice production and sustained draining of fresh water from the Arctic Ocean in response to winds were the key contributors to the

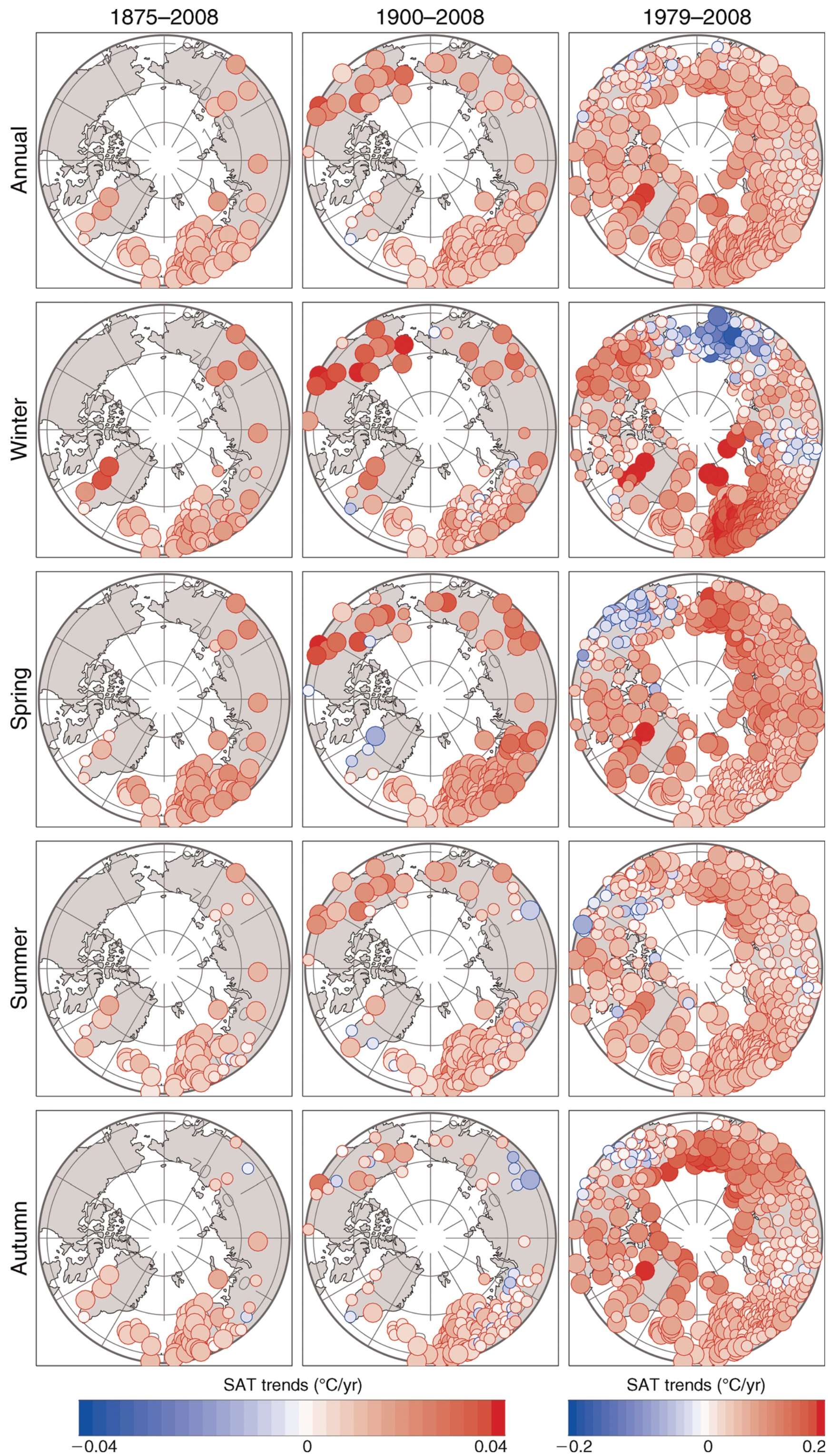

FIG. 3. Maps of annual and seasonal SAT trends based on meteorological station data. Statistically significant (95% confidence level) trends are marked by larger circles (from Bekryaev et al. 2010).

salinification of the upper Arctic Ocean during the 1980–1990s. The interpretation that the 2007 state of observed, anomalous FWC in the Canada Basin was due at least in part to peculiarities of ice–upper-ocean interactions, was confirmed through analysis of chemical tracers (Yamamoto-Kawai et al. 2009). This suggests that the Canada Basin is in transition from an ice-production region to an ice-melting region: a remarkable change in status for the polar basin in response to global warming.

Upper Arctic Ocean circulation is closely linked to atmospheric wind patterns. According to Morison et al. (2007), who analyzed data from a deep-sea pressure gauge and NASA's satellite-based Gravity Recovery and Climate Experiment during 2002–2006, the weakening of the Arctic Oscillation in the 2000s resulted in freshening of the upper Arctic Ocean near the North Pole, modifying the surface height and changing the oceanic circulation pattern. Morison et al. (2007) concluded that in the 2000s, circulation changed from the anticyclonic pattern that prevailed in the 1990s to the cyclonic pattern that was dominant prior to 1990. They also noted that this change in pattern is probably attributable not to climate change but to natural climate variability.

Anomalous Arctic atmospheric circulation contributes substantially to local FWC changes and freshwater exports (e.g., McPhee et al. 2009, Proshutinsky et al. 2009, Giles et al. 2012, Morison et al. 2012). For example, Timmermans et al. (2011) demonstrated the release of fresh water from the Beaufort Gyre and the freshening of the western Eurasian Basin in 2010 with potential consequences for freshwater export in subpolar basins. The importance of the export of freshwater anomalies from the Arctic Ocean into the Greenland Sea and further into North Atlantic regions was also stressed by Peterson et al. (2006). Further, Polyakov et al. (2008) found an out-of-phase relationship between the central Arctic Ocean and Greenland Sea FWC changes since the 1920s–1930s, which suggests that the strength of the export of Arctic ice and water controls the supply of Arctic freshwater to subpolar basins, while the intensity of Arctic Ocean FWC anomalies is of less importance.

Sea ice: observed changes and key pathways

Since 1979, Arctic sea ice extent at the end of the melt season (September) has declined at a rate of more than 11% per decade, and there is evidence that the rate of decline has accelerated over the last decade. Ice extent for September 2012 was the lowest in the satellite record for the month (see Plate 1), while 2007, 2011, and 2008 were the second, third, and fourth lowest, respectively. Further, September 2010 was the fifth lowest, and September 2009 was the sixth lowest (NSIDC 2011). Therefore, the lowest six end-of-summer extents have occurred in the past six years, and September ice extent has been more than two standard deviations below the long-term climatic mean in each of these six years. The winter ice extent has also been declining, albeit at a slower rate. Sea ice cover is also thinning (e.g., Rothrock et al. 1999, Kwok and Rothrock 2009). According to Kwok et al. (2009), the Arctic Ocean has lost 40% of its multiyear ice in the last five years. At the end of summer 2010, less than 15% of the ice remaining in the Arctic was more than two years old, compared to 50–60% during the 1980s (NSIDC 2010). The oldest sea ice remaining in the Arctic (at least five years old) has decreased to less than 60 thousand km^2, compared to 2 million km^2 during the 1980s, thus constituting only 3% of the Arctic ice area.

Over the past decades, the ice of the Arctic Ocean has become not only younger (Nghiem et al. 2007, Belchansky et al. 2008) but also more mobile, particularly in terms of increased velocities when leaving the Arctic Ocean through Fram Strait. For example, Kwok (2009) found an increase of the sea level pressure gradient across Fram Strait, which is generally associated with increased winds and ice drift speed. At the same time, Kwok (2009) noticed that the decrease in concentration of exported ice has more than offset an increase in the sea level pressure gradient across Fram Strait, resulting in "no statistically significant trend in the Fram Strait area flux" (Kwok 2009:2438). Within the Arctic Ocean, Ogi et al. (2008) showed that wind-driven sea-ice drift during summer has played an important role in regulating annual minimum Arctic sea ice extent during the past few decades, especially in 2007. Ogi et al. (2010) found that the combined effect of winter and summer forcing explains about one-third of the downward trend of sea ice extent from 1979 to 2009.

This observed Arctic sea ice reduction results from a complex interplay between the dynamics and thermodynamics of the atmosphere, sea ice, and ocean. The partitioning between these governing forcings remains a central research question (e.g., Lindsay et al. 2009, Serreze et al. 2009, Polyakov et al. 2012). Dynamical mechanisms such as wind-driven ice compression toward the western Arctic, enhanced transport of ice by the Transpolar Drift through Fram Strait, and ice export to Baffin Bay were the major causes for the dramatic summer 2007 ice loss (e.g., Nghiem et al. 2007). At longer time scales, however, the balance of forces may be different. For example, using observations and modeling, Polyakov et al. (2010) concluded that in the eastern Arctic over several recent decades, the added ice-bottom melting due to oceanic heat flux was comparable to the added surface melting due to atmospheric thermodynamic forcing. These two forcings were responsible for ~60% of the total ice thickness loss, and they may have preconditioned the ice cover, leaving it vulnerable to anomalous wind events, such as those of 2007 (Serreze and Francis 2006, Stroeve et al. 2012).

Over the most recent decade, the polar basins have begun to show reaction to the diminished Arctic ice. Observations suggest an almost synchronous reduction in the warm anomaly over the Eurasian Basin (Polyakov et al. 2011), raising the possibility that local atmo-

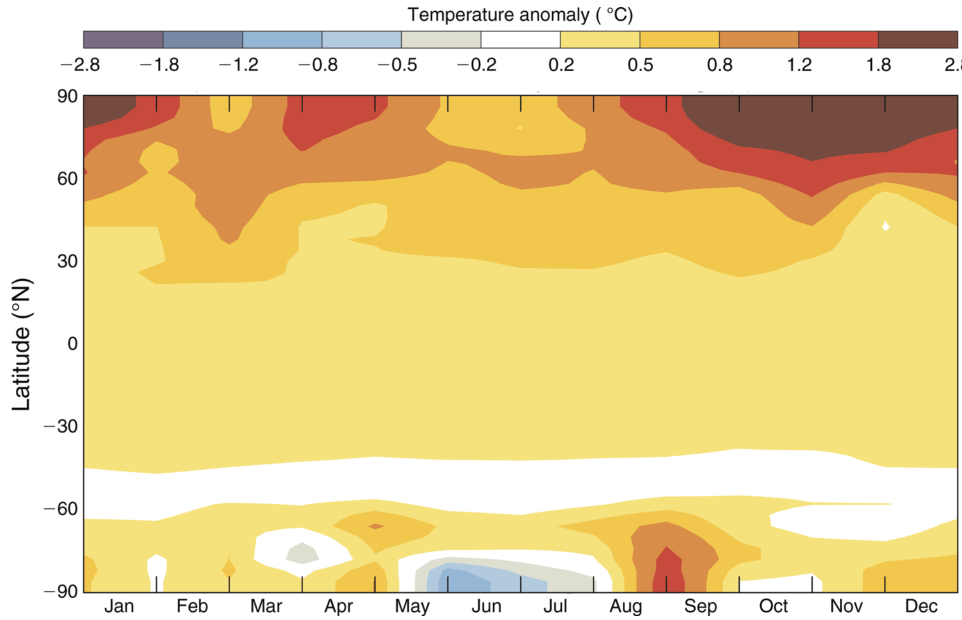

Fig. 4. Latitudinal vs. monthly mean temperature anomalies for 2001–2010, relative to corresponding means for 1951–2000. The globally averaged temperature anomaly for 2001–2010 is 0.45°C. Data are from NASA Goddard Institute for Space Studies, (http://data.giss.nasa.gov/gistemp/seas_cycle.html).

sphere–ocean and shelf-basin interactions influenced by anomalous openings in ice cover likely play an important role in ventilating the ocean's interior (e.g., Ivanov and Watanabe 2013). Another example of oceanic response to ice loss is a four-fold amplification of the solar, semi-diurnal, tidal current found under ice-free conditions in the eastern Eurasian Basin (Pnyushkov and Polyakov 2012). Thus, the predicted erosion of Arctic ice may further expose vast abyssal areas that rim the polar basins to an amplification of tides, enhancing vertical mixing rates and heat transfer from warm intermediate layers of the Arctic Ocean to the surface, and eventually to the ice cover. This potential positive feedback mechanism may further amplify the rate of Arctic ice decay and may also have important implications for Arctic marine ecosystems (*Feedback pathways of Arctic marine ecosystems*).

As sea ice is lost, the surface energy budget and associated fluxes to the atmosphere change to the extent that impacts on the atmosphere become possible, if not likely. The discussion above has already shown that Arctic SATs have been impacted by the loss of sea ice and snow. Beyond local impacts, there is evidence that non-local signatures of the loss of sea ice are emerging. Bhatt et al. (2010) showed that sea-ice variations of the past two decades have been significantly correlated with summer warmth in the tundra regions, particularly over North America, and with vegetation photosynthesis

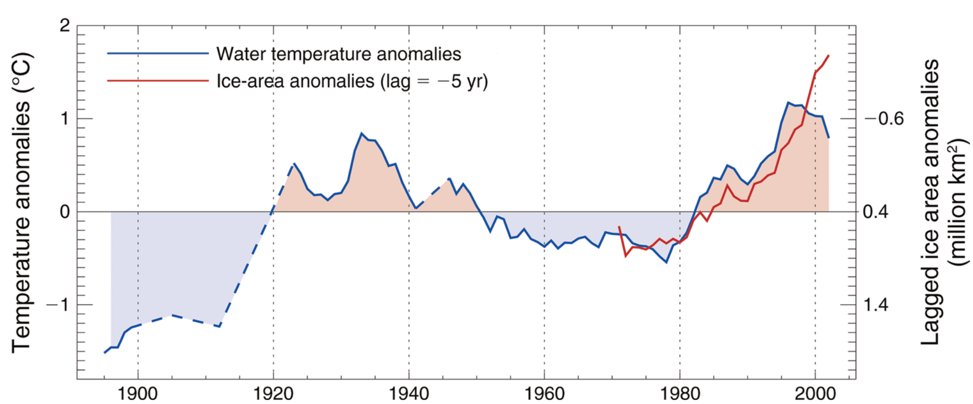

Fig. 5. Seven-year running mean normalized (dimensionless) intermediate Atlantic Water (AW) core temperature anomalies (dashed segments represent gaps in the record) and September Arctic ice-area anomalies (reverse vertical axis is used, lagged by five years; courtesy NSIDC and IJIS). AW core temperature is defined by its maximum, found at ~200–250 m depth range in the Eurasian Basin and deeper in the Canadian Basin (from Polyakov et al. 2010).

capacity. This evidence is consistent with the model-derived linkage between sea-ice loss and permafrost thaw in the Arctic near-shore terrestrial regions (Lawrence et al. 2008).

There are recent observational indications that the loss of sea ice has exerted a detectable influence on large-scale atmospheric circulation. Francis et al. (2009*a*) showed a linkage between summer sea-ice conditions and the atmospheric circulation of the following winter, while Overland and Wang (2010) have pointed to a feedback of sea-ice loss into the atmospheric circulation of autumn and early winter. In the latter study, reduced sea ice was shown to have contributed to an increase in upper air pressures (an air column expands vertically when heated from below) and a corresponding reduction in westerlies in recent years. The intriguing possibility is a linkage between these reduced westerlies and the anomalously negative phase of the Arctic Oscillation during the early period of the winters of 2009–2010 and 2010–2011. These extreme, negative Arctic Oscillation phases brought exceptional cold and snow to Europe and the eastern United States during these two periods. More recently, Liu et al. (2012) and Francis and Vavrus (2012) have provided observational and modeling support for the notion that reduced Arctic sea ice cover in autumn affects wintertime atmospheric circulation, particularly the frequency of blocking events that in turn lead to persistence or slower propagation of anomalous temperatures in middle and high latitudes. Liu et al. also point to a linkage between reduced sea ice and increased winter snowfall; enhanced evaporation from expanded open-water areas in the Arctic makes such a linkage physically plausible in autumn and early winter. While natural variability cannot be dismissed as the explanation of these anomalies in recent winters, the physical consistency of sea-ice loss and early-winter reductions in zonal flow implies that the loss of sea ice may already be influencing large-scale atmospheric circulation. This represents an illustration of a pathway linking sea ice and the atmosphere, though this and other atmosphere-ice pathways do not yet comprise a closed feedback loop. Nevertheless, this example of Arctic ice–ocean–atmosphere coupling has consequences for middle latitudes as well as the Arctic.

Trajectories into the future

The trajectory of the Arctic's coupled sea ice, ocean, and atmosphere has emerged as a focus of research, especially in the years since the extreme sea ice retreat of 2007 and the widespread loss of thick multiyear ice. While there is still some debate about the relative importance of drivers (atmosphere vs. ocean, short-term forcing vs. preconditioning), there is general consensus that the continued loss of sea ice is likely, as there are no indications that the Arctic atmosphere and Arctic Ocean interior will become cooler soon (Table 1). In particular, there is little reason to expect that the pre-conditioning effect of Arctic Ocean Atlantic Water on the polar ice cap will subside in the near term, although there may be short-term periods (decadal-scale and shorter) during which this Atlantic Water cools. There are also finite probabilities (about 30%, according to experiments with the CCSM4 model) that any particular decade could see an increase in sea ice coverage from the preceding decade (Kay et al. 2011). However, aside from natural variability, there are no signs of any systematic slowing of the tendency for sea ice and the upper ocean to become more mobile in the Arctic (via reduced ice area and the associated increased impact of winds on ice and upper-ocean circulation, enhanced tides, and so on). Shelf–basin interactions are likely to become increasingly active, altering rates of heat exchange, especially near the shelf break, with implications for marine ecosystems in the shelf seas of the Arctic. In addition, the eastern Arctic Ocean's tendency to become an ice-melt area (compared to its history as an ice-generation area) should continue in the absence of any countering tendencies (Table 1). Over the long term, the main questions concerning these tendencies are not as much about future direction as they are about rates of change. Attempts to narrow uncertainties in model-projected trends indicate that the most likely timeframe for the development of a seasonally ice-free Arctic Ocean is mid-century, perhaps even sooner. Since the projected decrease in ice extent is much less in winter than in summer, a seasonal (first-year) regime of sea ice will prevail. The cold-season ice cover will be thinner, more mobile and more deformable than in the past. However, one must acknowledge the possibility of surprises, whether the result of natural variability (e.g., wind anomalies such as those of 2007, or even the reverse pattern) or processes whose importance have not yet been recognized (Kattsov et al. 2010, Stroeve et al. 2011).

The trend toward a thinner, seasonal ice cover and warmer upper ocean will likely increasingly influence the atmosphere through heating, especially in winter and autumn. Temperature inversions will become weaker and less frequent, and precipitation will increase as the open-water season lengthens. Increased precipitation, together with increased discharges from rivers and ice sheets, will contribute to a freshening of the Arctic's subpolar seas. Intensification of this freshwater pathway may contribute to feedbacks in both the terrestrial and marine components of the Arctic system (*Feedback pathways of Arctic land ecosystems* and *Feedback pathways of Arctic marine ecosystems*).

The enhanced atmospheric heating arising from the loss of sea ice raises the likelihood of an effect on large-scale atmospheric circulation, with attendant impacts on middle latitudes. As noted in *Sea ice: observed changes and key pathways*, an ice-driven tendency for increased atmospheric blocking will mean greater variability of associated circulation-driven anomalies in the middle latitudes (Francis and Vavrus 2012). In this respect, the extreme variability of the past three winters (2009–2010, 2010–2011, and 2011–2012) in the United States may well be a preview of the future.

There are still fundamental questions about whether poleward heat transports will increase or decrease if sea-ice loss contributes to a polar-amplified warming of the Northern Hemisphere. Pole-to-equator temperature gradients should weaken under polar amplification, but any trend toward a more meridional atmospheric circulation would favor an increase of heat transport into the Arctic. We are just now entering the phase of climate research in which component interactions may be quantified sufficiently to address feedbacks in the atmosphere–ice–ocean system. The trajectory of the Arctic climate system over the coming decade, together with enhanced research on the dynamics of climate change, will likely elucidate the interactions and feedbacks that arise from sea ice loss.

Finally, atmospheric and oceanic warming and sea-ice loss in the Arctic are likely to increase atmospheric water vapor content and to change Arctic cloudiness (see also Table 1). Water vapor is an important greenhouse gas, and clouds profoundly affect the surface radiative budget. Serreze et al. (2012) have shown that sea-ice loss is already increasing the Arctic's atmospheric water vapor content. While there are indications that Arctic cloudiness is increasing in some seasons (Wang and Key 2005), there is little evidence of systematic changes in Arctic cloudiness in other seasons. Even the impacts of clouds are difficult to quantify in a broader framework of climate change, as cloud-radiative forcing of the Arctic surface is generally positive in winter and negative in summer. One must therefore acknowledge that Arctic clouds and water vapor will shape future Arctic change, but that the key processes and feedbacks by which they will do so are inadequately understood at the present time.

Key Pathways Involving the Greenland Ice Sheet, Glaciers, and Ice Caps

The Greenland ice sheet: changes and connections

The Greenland ice sheet (GrIS) is the largest reservoir of permanent snow and ice in the Arctic (~7 m sea-level equivalent [SLE]), and it is highly sensitive to ongoing climatic variations and changes (e.g., Hanna et al. 2012). Since 1979, satellite observational studies (Mote 2007, Tedesco 2007, Steffen et al. 2008, Tedesco et al. 2011) have demonstrated that the GrIS melt extent has increased (Table 1), and that the extent is highly sensitive to climate changes, with record melt extent in 2007 and again in 2010. Further, GrIS melt from 1960–1972 was simulated by Mernild et al. (2011*b*), who indicated an average decrease in melt extent of 6%; 1973–2010, however, showed an average increase in melt of 13%, again, with record melt extents in 2007 and 2010. This trend in simulated melt extent since 1972 also indicated that the melt extent in 2010 averaged twice that of the early 1970s, with maximum melt in 2010 extending to over 52% of the GrIS area. Furthermore, the melt period increased by an average of two d/yr since 1972, indicating an extended melting period for the GrIS of about 70–40 days and 30 days in the spring and autumn, respectively.

The changes in the GrIS melting regime (extent and duration) can produce substantial differences in surface albedo and energy and moisture balances, and may establish a positive feedback mechanism, since wet (i.e., below the dry-snow line) snow absorbs up to three times as much incident solar energy as dry (above the dry-snow line) snow (Steffen 1995). An altered melt regime can influence the ice sheet's surface mass balance (SMB), including its runoff, ice-surface velocity, and subglacial sliding processes. One would expect that increasing annual ablation would cause increasing mean annual ice-flow velocity, as the increasing volume of lubricating surface meltwater reaches the ice-bedrock interface (Joughin et al. 2008). However, for a land-terminating sector of the GrIS, it has been shown that a higher-average annual ablation does not necessarily lead to an increase in annual ice velocities (van der Wal et al. 2008). Mechanisms that link climate, surface hydrology, internal drainage, and ice dynamics are poorly understood, and numerical ice-sheet models do not simulate these changes realistically (Nick et al. 2009).

From 1960 to 2010, the GrIS surface hydrological conditions, net accumulation (precipitation minus evaporation and sublimation), runoff, and SMB (Fettweis et al. 2008, Hanna et al. 2008*a*, Ettema et al. 2009, Mernild and Liston 2012), were influenced climatically first by decreasing air temperature (1960 to mid-1980s) and then by increasing air temperature and net precipitation (mid-1980s to present; Cappelen 2011), and the subsequent increasing surface runoff, leading to enhanced SMB loss. Overall, for the GrIS, the amount of runoff to adjacent seas explains about half of the recent annual GrIS mass loss (Rignot and Kanagaratnam 2006, van den Broeke et al. 2009), with iceberg calving generating the other half.

For the GrIS, subtracting the average surface runoff from the net precipitation yielded a surface mass gain, with an average annual GrIS SMB of 156 ± 82 km^3/yr (1960–2010; mean ± SD), varying from 220 ± 86 km^3/yr in 1970–1979 to 1986 ± 72 km^3/yr in 2000–2010 (Mernild and Liston 2012). Mernild et al. (2010) further projected, on average, a GrIS SMB mass loss after 2040, based on the IPCC A1B scenarios: values in the range of projections by Fettweis et al. (2008). While the GrIS surface was gaining mass (1960–2010; with a positive SMB) and closely following mean temperature fluctuations on average, the overall mass balance for the GrIS was close to equilibrium during the relatively cold 1970s and 1980s and lost mass rapidly as the climate warmed in the 1990s and 2000s, with no indications of deceleration (Rignot et al. 2008). Since 2003–2009, the average annual GrIS mass loss rate was −250 km^3/yr (equal to −250 Gt/yr); in particular, from April 2010 to April 2011, the ice sheet cumulative loss rate was −430

Gt/yr: 70% larger than the 2003–2009 average annual loss rate (Wahr 2011). This represents a GrIS loss rate equivalent to a eustatic sea-level rise contribution of 1.1 mm SLE/yr, compared to a mean global sea level rise of 3.3 ± 0.4 mm SLE/yr from 1993 through 2009 (Nerem et al. 2010).

In addition to the sea level contribution, the increase in GrIS runoff (in net amount and duration), and in iceberg calving (in mass loss) are important for ocean density changes and the thermohaline circulation (Rahmstorf et al. 2005), specifically to the Atlantic Meridional Overturning Circulation (AMOC) and its impact on the climate system (Bryden et al. 2005). Since the early 1960s, observed time series of salinity around Greenland (e.g., from the East and West Irminger Seas, Danmark Strait, Labrador Sea, Faroe-Shetland Channel, and Fram Strait) indicate overall decreases in average salinities (Dickson et al. 2003). The AMOC may be sensitive to changes in salinity based on GrIS runoff and calving; freshening of surface waters in the northern North Atlantic inhibits deep convection that feeds the deep southward branch of the AMOC (e.g., Rahmstorf 1995). The AMOC carries warm upper waters into far-northern latitudes and returns cold deep waters southward to the southern hemisphere. This heat transport contributes substantially to the climate of continental Europe and the Northern North Atlantic region (Hanna et al. 2008*b*), for example, and any slowdown in the circulation due to increasing terrestrial freshwater flux could have implications for climate change (Bryden et al. 2005). This pathway between the GrIS and North Atlantic/European climate could be a key linkage in a broader climate feedback via ocean circulation, though the extent to which GrIS-induced climate change would, in turn, alter the GrIS is not known.

Not only has the GrIS lost mass, it has also faced area recession (Kargel et al. 2011). The land- and marine-terminating outlet glaciers on the periphery of the GrIS have undergone rapid area changes over the last decades (Howat and Eddy 2011, Mernild et al. 2012), during which a transition occurred from stability and small fluctuations in marine-terminating glacier frontal positions (1972–1985), to moderate widespread recession in the southeast and western parts of the GrIS (1985–2000), followed by an accelerated net recession in all regions of the ice sheet (2000–2010) (Howat and Eddy 2011). For 2000–2010, the mean net recession rate for frontal positions of GrIS marine-terminating outlet glaciers was 0.11 km/yr, for 210 outlet glaciers (Howat and Eddy 2011). Box and Decker (2011) measured area change at 39 of the widest GrIS marine-terminating glaciers based on Moderate Resolution Imaging Spectroradiometer (MODIS), between 2000 and 2010. Overall, for these 39 glaciers, they exposed a cumulative net area of 1368 km^2. The recession in GrIS cover for the last decades for both land- and marine-terminating margins caused changes in the surface albedo and subsequent changes in the energy balance. This climate/ice-sheet pathway may establish a positive feedback mechanism and an increase in GrIS melting, for example, due to a higher degree of surface energy absorption.

Glacier and ice cap changes and connections (outside the Greenland ice sheet)

Half of the estimated global glacier and ice cap (GIC) surface area, and two-thirds of the GIC volume, are located in the circumpolar Arctic region (Radić and Hock 2010). This makes studies of SMB and volume changes in northern latitude GIC essential because of their contribution to global sea-level rise. However, few GIC SMB observations exist. SMB observations from around 25 GICs in the circumpolar Arctic (Alaska, Arctic Canada, Greenland, Iceland, and Svalbard) have been published for 2001–2010 (Haeberli et al. 2009), representing a minor fraction of the Arctic's several thousand GICs. Further, very few GIC have uninterrupted annual records covering at least two or more decades. Even though GICs account for less than 1% of all water on Earth bound in glacier ice, their increasing retreat and mass loss due to iceberg calving and runoff may have dominated the glacial component of the global eustatic sea-level rise during the past century (Leclercq et al. 2011, Radić and Hock 2011). At present, GIC mass losses are raising the mean global sea level by approximately 1.0 mm SLE/yr (Table 1). It has been suggested that GIC SMB losses are expected to dominate the cryospheric contribution to sea-level rise between now and 2100 (Cogley 2012), and that the GIC-derived sea-level rise will continue to increase in the future (Dyurgerov and Meier 2005, Meier et al. 2007). Recent analyses based on direct and geodetic measurements suggest that GIC mass loss is raising sea levels by about 1 mm/yr (Table 1). This is about one-third of the total rate of sea-level rise inferred from satellite altimetry, with ocean thermal expansion and ice-sheet mass loss accounting for most of the remainder (Cazenave and Llovel 2010). The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4; IPCC 2007) projected 21st-century global sea-level rise due to GIC wastage to be in the range of 0.07–0.17 m SLE: approximately one-third of the total 21st-century sea-level rise. Radić and Hock (2011) projected that sea-level rise from GIC wastage by 2100 will amount to 0.124 ± 0.037 m (mean ± SD), with the largest contribution from Arctic circumpolar GICs in Arctic Canada and Alaska. Radić and Hock (2011) GIC sea-level projections are the most detailed to date (Cogley 2012).

GIC mass balance observations for the last decades have shown an overall increase in mass loss (Table 1; Dyurgerov 2010, Cogley 2012), heading toward more negative annual SMB during the first half (2001–2005) of the first decade of the 21st century (Zemp et al. 2011). These negative balances of the early 2000s have not been sustained during the most recent five-year period (2006–2010), in which GIC losses have been more moderate, though still large (Cogley 2012). For example, Gardner et al. (2011) showed that Canadian Arctic Archipelago GIC had recently lost 61 ± 7 Gt/yr of ice for the period 2004–2009. In East Greenland (peripheral to the GrIS), GICs have also lost mass and faced area recession. A study by Mernild et al. (2012), based on satellite observations of 35 GICs in East Greenland, indicated a mean areal recession of around 1% per year, indicating that GICs overall have lost 27% ± 24% (mean ± SD) of their area since 1986. Further, five GICs have melted away, and GICs on average have faced a volume recession of one-third since 1986 (Mernild et al. 2013*a*). Specific GICs in East Greenland might be significantly out of equilibrium with the present-day climate, and will likely lose at least 70% of their current area and 80% of its volume, even in the absence of further climate changes. Globally, GICs will shed around one-quarter of their present-day volume if they are to reach a size in equilibrium with the present-day climate (Mernild et al. 2013*b*), indicating that Greenland GIC area and volume adjustment lag behind climatic forcing (Cogley 2012). Temperature records from coastal stations in East Greenland suggest that recent GIC area and volume losses are out of equilibrium and not merely a local phenomenon, but indicative of glacier changes in the broader region (Mernild et al. 2011*a*).

Key Pathways in Hydrology and Permafrost

Feedback pathways from permafrost degradation

Perhaps the first incontrovertible evidence of late-20th-century climate warming was published by Lachenbruch and Marshall (1986), in which borehole measurements showed that permafrost on the North Slope of Alaska was in a warming phase. In recent years, permafrost warming and thawing has been documented throughout the circumpolar Arctic (Table 1; Hinzman et al. 2008, Osterkamp 2008, Romanovsky et al. 2010*a*, *b*, Smith et al. 2005*b*, 2010), and permafrost warming has significantly accelerated along many parts of the Arctic coast during the last five years (Romanovsky et al. 2010*b*). Ice wedges, which form in permafrost and only survive as long as permafrost remains frozen, provide good information about the age of the permafrost and previous climates under which the permafrost has existed. In northern and central Alaska, permafrost has been radiometrically dated between 14 000 and 32 000 years old, (Péwé 1973) and ice wedges throughout the Arctic are assumed to be Pleistocene to Holocene in age (Brown and Péwé 1973). Over these long time periods, the permafrost grew to great depths, up to 600 m in northern Alaska and more than 1000 m in Siberia (Brown et al. 1997); therefore, warming and complete thawing of deep and cold permafrost will be a long, slow process. However, warmer permafrost (>−1°C) in areas of sporadic or discontinuous permafrost is particularly vulnerable and is already degrading under a warming climate (Jorgenson et al. 2001). Normally, heat flow in permafrost occurs via conduction, which is limited by thermal conductivity and small thermal gradients (Kane et al. 2001). Additionally, latent heat requirements to thaw permafrost and melt the massive ice can be substantial, greatly slowing the natural thaw of permafrost (Grosse et al. 2007). However, permafrost may thaw rapidly in situations influenced by running water and thermal and mechanical erosion.

In general, the active layer (the layer of soil above permafrost that annually experiences thawing and refreezing) ranges from about 20 cm to 1 m in thickness, depending upon substrate, moisture, and exposure. In the Alaskan Kuparuk basin, on average, the active layer thickness is about 40–60 cm deep (Hinzman et al. 1998), greater in well-drained sites and shallower in poorly drained sites. Modeling studies usually predict deeper active layers with a warming climate (Schaefer et al. 2012), and recent increases in active layer thickness have been observed in Subarctic Scandinavia (Åkerman and Johansson 2008), Greenland (Christiansen and Humlum 2008), and portions of the Russian Arctic (Mazhitova et al. 2008), but despite reliable observations of increasing air and permafrost temperatures, consistent increases in active layer thickness have not been documented in many parts of the Arctic (data *available online*).[6] This puzzling lack of apparent response in the active layer may be due to the fact that most atmospheric warming has occurred in the winter, while active layer thawing is primarily responsive to summer temperatures. The lack may also be due, however, to thawing of excess ice at the base of the active layer as summer warming increases (Overduin and Kane 2006). Overduin and Kane (2006) have shown that over a three-year period, no increase in active-layer thickness was observed for a site on the North Slope of Alaska from relative depth measurements at summer's end, despite successive increases in air temperature. However, more detailed analyses showed that areal ground surface subsidence rates of 2–5 cm/yr were due to the melting of ice at the base of the active layer.

Increases in active-layer thickness at some locations have led to subtle but predictable ecosystem responses, such as vegetation changes. Sturm et al. (2001) described the increase in shrubs in the present tundra environment on the North Slope of Alaska, which has implications for trapping snow, evapotranspiration, soil moisture, and runoff, potentially. More shrubs will lead to deeper snowpack, which in combination with predicted greater winter precipitation (Rawlins et al. 2010) will in turn provide greater insulation for the ground from the cold winter air temperatures, resulting in a warming of permafrost (Stieglitz et al. 2003). A warmer climate should result in a deeper active layer initially; however, it is possible that shrubs will provide ground shading and

[6] http://www.udel.edu/Geography/calm/data/north.html

thus lower ground-surface temperatures. Different plant communities have different soil water demands, and variations in microtopography and subsurface drainage properties will also introduce heterogeneity in soil moisture content. Drier soils typically have deeper active layers, as less energy is needed for phase changes during thawing. As the active layer thickens, there is greater storage capacity for soil moisture, and greater lags and longer recessions are introduced into hydrologic response times to precipitation (Bolton et al. 2004). When the frozen ground is very close to the surface, the stream and river discharge peaks are higher and base flow is lower.

Winter precipitation can also play a conflicting role in regulating active ground temperatures and active layer thickness. Except for extremely dense snowpack, such as wind-packed or rain-affected snow, most snow conditions have relatively low thermal conductivity (Sturm et al. 2002), offering some insulation to the permafrost from extremely cold winter air temperatures. Increases in permafrost, perhaps in response to the equally well-known increases in winter air temperature, are widely documented, but warming permafrost may also be a response to the highly variable, though generally increasing trends in snowpack thickness (Table 1; Pavlov and Moskalenko 2002, Troy et al. 2012). The effects of thicker snowpack on active-layer thickness are more difficult to predict. Although small increases in active layer thickness following winters with greater snowpack have been documented (Frauenfeld et al. 2004), this is somewhat confounded with the fact that winters with greater snowfall tend to be warmer and tend to experience delayed snowmelt and delayed onset of ground thawing (Kane et al. 2000).With increased warming and thawing, the active layer will continue to increase in thickness and will eventually reach a point where it can no longer refreeze completely during the following winter, leaving a layer of unfrozen soil between permafrost and seasonally frozen ground. This layer of unfrozen soil, called a talik, can quickly exert dramatic effects upon surface water and energy balances (Hinzman et al. 2003*a*). The talik may permit the soil to continue to drain throughout the winter, creating drier conditions in the soil column, with consequent effects to the local ecosystem. The maximum depth to which soil may freeze is usually in the order of 1–2 m, depending upon moisture content, soil type, snow cover, and the thickness of surface organic layers. If summer thawing consistently exceeds this depth and a talik is established and maintained, then the surface may begin to exhibit conditions consistent with an area that is permafrost free. Formation of a talik essentially crosses a threshold, in which soils may go from cold and wet to warmer and drier, with important implications upon local vegetation, hydrology, and surface energy balance.

Extensive surveys conducted throughout the circumpolar Arctic have documented the acceleration of coastal erosion rates (Forbes 2011), which has introduced tremendous amounts of long-frozen carbon directly into the marine ecosystem (Table 1). Warmer permafrost and sea-water temperatures, combined with less extensive sea ice, have increased the vulnerability of Arctic coasts to erosion. In Alaska, rates of coastal retreat were most commonly in the 1–2 m/yr range, but these rates have varied up to 10–30 m/yr in some locations over the past decade (Jones et al. 2009). Erosion of the Laptev Sea coast at rates of $\sim$2.5 m/yr has delivered more sediment and carbon into the sea than the Lena River (Rachold et al. 2000). Trends of decreasing sea ice and increased open-water fetch are expected to result in higher wave energy, and when combined with warming air and ground-surface temperatures in summer, will continue to increase seasonal thaw and accelerate coastal retreat along large parts of the circum-Arctic coast. As on temperate coasts, waves, currents, and water levels are major forcing parameters; however, in the Far North, sea-surface temperature and salinity are also very important, in that they contribute to thawing of ice-bonded sediment in the beaches and backshore. Extensive coastal erosion, most significantly in Canada, Alaska, and Siberia, represents a potentially important pathway between the terrestrial and marine Arctic. Feedbacks may arise if the coastal marine environment changes sufficiently to impact the atmosphere or coastal land areas. Decomposition of the tremendous stores of organic carbon frozen in permafrost has been identified as one of the most significant potential feedbacks from terrestrial regions (Schuur et al. 2008). The initial ecosystem response to permafrost degradation is increased decomposition of organic matter, with the consequent release of soil nutrients and enhanced plant growth due to this fertilization effect (Schuur et al. 2009), resulting in greater carbon sequestration and a negative feedback to atmospheric warming. However, with continued permafrost degradation, carbon sequestration due to enhanced vegetation growth is eventually surpassed by carbon emissions to the atmosphere, and the net effect is a positive climatic feedback. Carbon feedbacks are presented in more detail in *Feedback pathways of Arctic land ecosystems: Observed changes: Exchange of radiatively active gases with the atmosphere*.

Feedback pathways from hydrology changes

Due to typically low rainfall rates and the frequent occurrence of wind that often accompanies rain and snow, precipitation in the Arctic is both difficult to measure and complex to predict (Yang et al. 2005). To this end, Houghton et al. (2001: Fig. 2.25ii) suggested that a 1% increase in precipitation per decade was probable over the last century, and Walsh et al. (2012) have reported that recent pan-Arctic annual precipitation has increased by about 5% since the 1950s (Table 1). Further, Rawlins et al. (2010) compiled extensive analyses of gridded precipitation data in order to assess trends in Arctic terrestrial regions. This analysis of pan-

Arctic (excluding Greenland) annual precipitation revealed trend-line slopes ranging in magnitude from –0.03 to 0.79 mm/yr^2. A significant positive trend of 0.21 mm/yr^2 was determined from the Climatic Research Unit (CRU) version 3.0 data set for the period 1950–2006 (*available online*).[7] Similarly, compiled analyses of gridded evapotranspiration yielded exclusively positive trend-slopes, ranging from 0.11 to 0.40 mm/yr^2 (Table 1). Data ranges varied from 1950–1999 to 1989–2005, with latter data periods demonstrating greater trend-line slopes. Most of the climate stations reported by Hinzman et al. (2005) showed an increase in annual precipitation over the length of their record (from the late 19th century to more recently), though summer surface-water balance (precipitation minus potential evapotranspiration, P-PET) measured in Alaskan North Slope villages had decreased since 1960 (Oechel et al. 2000), consistent with longer and warmer summers (hence greater ET). This seasonal distribution of precipitation is important to consider, as Arctic winter precipitation is projected to increase with continuing climate change (Serreze et al. 2000, Skific et al. 2009). However, Liston and Hiemstra (2011) conducted an extensive modeling study coupled with analysis of remotely sensed data of the pan-Arctic domain for the satellite era (1979–2009) and discovered that maximum winter snow-water equivalent has decreased, snow-cover onset is later, snow-free date in spring is earlier, and snow-cover duration has decreased, although positive and negative regional trends are distributed throughout the study domain. Brown et al. (2010) confirmed decreases in snow-cover duration and further noted that observed reductions in June snow-cover extent were of the same magnitude as reductions in June sea-ice extent, and that both were significantly correlated to air temperature changes. Snow-cover extent has declined significantly during spring across both North America and Eurasia, albeit with lesser declines during winter and some increases during fall (Déry and Brown 2007). The overall decline appears to be driven by and may have contributed to polar amplification of springtime warming; this amplification provides further evidence of surface-atmosphere coupling in the Arctic (Ghatak et al. 2010). There is also evidence that snowpack structure is changing as a result of more frequent mid-winter melt events and increased rain on snow (Brown et al. 2012). This could represent an important pathway to the ecological system through restricting browsing by herbivores and encouraging the propagation of select plant species.

The net result of changes in precipitation minus evapotranspiration is more difficult to evaluate, as this would be reflected in changes in river discharge and changes in surface and subsurface storage. A number of studies have demonstrated substantial long-term changes in Arctic river discharge, primarily in Russia, but more recently also in North America (Table 1; White et al. 2007). Combined discharge from the six largest Russian Arctic rivers increased by about 7% over the 1936–1999 period (Peterson et al. 2002), with most of the increase occurring during the past few decades (Yang et al. 2002, McClelland et al. 2004). A trend analysis of Canadian rivers by Déry et al. (2009) demonstrated a dramatic 15.5% increase in annual flows over the recent past (1989–2007). These increasing trends occurred across most of northern Canada, with the exception of some rivers in the Hudson Bay area.

A number of studies have documented substantial changes in the size and number of surface water bodies in recent years. These studies include Roach et al. (2011), Jorgenson et al. (2006), and Yoshikawa and Hinzman (2003) in Alaska; Smol and Douglas (2007) in Canada; and Smith et al. (2005*a*) in Siberia. These studies have documented both increases and decreases in lake or pond size and number (Table 1). The reasons for such changes have been attributed to improved drainage associated with the degradation of near-surface or deeper permafrost as well as changes in water balance, usually as evaporation increased more than precipitation increased. Trochim et al. (2008) detected an increase in the number of small stream channels draining hillsides in a watershed in the Alaskan Arctic over the 1956–2007 period. An increase in the number of these water tracks will increase the drainage density and effectively allow water to flow from hillsides more efficiently, with a net result of enhanced drying.

The important hydrologic processes in terrestrial portions of the Arctic and Subarctic include snow and rainfall precipitation, redistribution of snow by wind, snowmelt, damming of snowmelt runoff by the snowpack, surface storage in small depressions and ponds, infiltration and storage in the substrate, evaporation, transpiration, and runoff (Woo 1986). In regions of discontinuous permafrost, groundwater also becomes an important factor. Surface moisture and surface temperature are the fundamental variables in coupled terrestrial and atmospheric processes, and these represent the dynamic linkages between atmospheric, hydrological, biological, and permafrost processes. Surface moisture and temperature appear repeatedly in many documented feedback mechanisms (Francis et al. 2009*b*), and thus exert a dominant influence upon many climatic and ecosystem processes. Surface temperature is the linchpin in energy fluxes, as it links atmospheric thermal gradients, forcing convective heat transfer, with the subsurface thermal gradients driving conductive heat transfer. Soil moisture exerts a strong influence upon energy fluxes directly (through controls on evaporative heat flux and phase change during thawing of permafrost) and indirectly (through effects on thermal conductivity).

In the Arctic, the primary control on local hydrological processes, with consequent effects on the

[7] http://badc.nerc.ac.uk/data/cru/

ecosystem and surface energy balance, is the presence or absence of permafrost, though the control itself is also influenced by the thickness of both the active layer and its own total underlying permafrost layer (Hinzman et al. 2003*c*). As permafrost becomes thinner or decreases in areal extent, the interactions of surface and sub-permafrost ground water processes become more important (Woo 1986). The inability of soil moisture to infiltrate to deeper groundwater zones due to ice-rich permafrost maintains very wet soils in Arctic regions (Hinzman et al. 2003*b*). However, in the slightly warmer regions of the Subarctic, permafrost is thinner or discontinuous. In permafrost-free areas, surface soils can be quite dry, as infiltration is not restricted, and this dryness may impact ecosystem dynamics, fire frequency, and latent and sensible heat fluxes. (Hinzman et al. 2006*a*, *b*). In warmer permafrost regions, sub-permafrost groundwater becomes more important, either by contributing groundwater into streamflow or allowing surface water to drain. A thermokarst occurs as a characteristic landform associated with degrading ice-rich permafrost as massive buried ice melts and the ground surface itself subside into resulting voids. Important and dynamic processes involved in thermokarst development include thaw, ponding, surface and subsurface drainage, surface subsidence, and related erosion. Other hydrologic processes impacted by degrading permafrost include increased winter stream flows, decreased summer peak flows, changes in stream water chemistry (Petrone et al. 2000), and other fluvial geomorphological processes. Hydrologic changes already witnessed include drying of thermokarst ponds (Table 1; Yoshikawa and Hinzman 2003), increasing importance of groundwater in the local water balance, and changes in the surface energy balance (*Feedback pathways of Arctic land ecosystems*).

Future trajectories of permafrost and hydrology

Although there is little question as to the direction of the trend, there is a great deal of uncertainty on the future rate of permafrost degradation in response to climate warming. One argument is that thaw is likely to proceed at a slow pace during the next century, as the thermal regime of permafrost is buffered by the overlying, protective layer of organic matter (Yi et al. 2007). Schaefer et al. (2012) have summarized permafrost modeling results from 13 different studies and have reported that areal loss of permafrost in the upper soil column range from 7% to 90% (averaging 44%) by the model year 2100, while the average active-layer thickness (in areas where permafrost persisted) shows an increase of 65 cm (Table 1). The simulations of Euskirchen et al. (2006), which considered the protective layer of organic matter, estimated that permafrost loss in the first 40 to 50 years of the 21st century would be minimal and primarily located in regions characterized by a discontinuous permafrost regime, at the transition zone between boreal and temperate forest ecosystems. These results compare well with modeling and observational studies conducted in Subarctic Alaska (Yoshikawa et al. 2002). As the active layer becomes thicker, both moisture storage capacity and the lag time of runoff increase. As permafrost becomes thinner, there may be more connections between surface and subsurface water and more infiltration to groundwater (Hinzman et al. 2003*c*). This has significant impacts on large and small scales. The timing of stream runoff will change, perhaps eventually reducing the percentage of continental runoff released during the summer and increasing the proportion of winter runoff (Yang et al. 2002).

Smith et al. (2007) demonstrated that glaciation history and the presence of permafrost appear most important to the existence of lakes. Furthermore, they noted that in a "permafrost-free" Arctic, the number of lakes could be reduced from ~192 000 to 103 000 (–46%), and their total inundation area reduced from ~560 000 to 325 000 km^2 (–42%). Although a permafrost-free Arctic may not be a realistic scenario, the undeniable result of permafrost thawing on continental scales will be net surface drying. On the local scale, in a scenario in which near-surface soils become permafrost-free, well-drained areas with low groundwater level (uplands) will become drier. At sites where the groundwater level is near the surface, soils may become wetter as permafrost degrades and the surface subsides.

Avis et al. (2011) examined surface drying using the UVic Earth System Model to simulate wetland loss under four forcing scenarios. Their simulations displayed an initial increase in wetland extent at very high latitudes (>70° N), which they attributed to increased precipitation. This increase in wetlands at high latitudes was offset by a loss of wetlands associated with permafrost degradation in warmer, discontinuous permafrost regions. Wetland loss occurred first at low latitudes and progressed poleward (Table 1), again indicating substantial surface drying following permafrost degradation. These results are consistent with the Smith et al. (2007) analysis described above and another Smith et al. (2005*a*) study, which documented an observed decrease in lakes in southern Siberia and an increase in lake area and number in northern Siberia.

Further, Serreze and Etringer (2003) have demonstrated that ~25% of high-latitude summer precipitation results from locally recycled evapotranspiration. A decrease in evapotranspiration fluxes (with all else equal) would therefore lead to a decrease in precipitation. If surface drying does follow permafrost degradation as projected, a reasonable assumption is that evaporation and evapotranspiration will decrease, resulting in a decrease in precipitation in summer.

Feedback Pathways of Arctic Land Ecosystems

Observed changes

Albedo and energy partitioning.—One important pathway through which land ecosystems of the Arctic influence the climate system is the changes in albedo

PLATE 1. International Arctic Research Center (IARC) researchers return to the same site in the Arctic Ocean, 78° N 150° W, every year since 2006 for research mooring recovery and redeployment; 2012 is the first year they observed no ice in this location. Photo credits: photos taken on board the CCGS *Louis S. St-Laurent* in 2006 (right) by Jenny Hutchings and 2012 (left) by Alice Orlich.

and energy partitioning of the ecosystem surface (i.e., the transfers of energy to the atmosphere as latent or sensible heat flux or as heat to the ground), which influence atmospheric warming or cooling. Specifically, changes in the length of the growing season can influence albedo and energy partitioning of terrestrial ecosystems in the Arctic. Several studies based on remote sensing indicate that growing seasons lengthened in high-latitude regions during the late 20th century (McDonald et al. 2004, McGuire et al. 2004), although there is evidence that growing season has not lengthened since 1997 (Piao et al. 2011). Collectively, these studies identify earlier thaw onset and later freeze-up, though the rate of change in growing season length differs in each study. Euskirchen et al. (2007) estimated that the regional changes in surface-energy exchange (i.e., reduced snow cover) associated with a longer growing season, in response to warming, were three times larger in the 1970–2000 warming period ($+0.9$ W$\cdot$m$^{-2}\cdot$decade^{-1}) than in the 1910–1940 warming period (Fig. 6). Tundra ecosystems, which have a high albedo contrast between summer and winter, were estimated to have much larger changes than forest ecosystems, which have a low seasonal contrast in albedo.

Changes in vegetation composition and distribution can affect albedo and energy partitioning of terrestrial ecosystems in the Arctic. Repeat aerial photography and remote sensing studies have documented an expansion of deciduous shrubs in tundra areas in northern Alaska during the second half of the 20th century (Tape et al. 2006). Shrub cover has been estimated to have increased in northern Alaska by about 16% in land area since 1950 (Tape et al. 2006), and shrub growth appears to be increasing throughout the tundra of much of the Arctic (Myers-Smith et al. 2011). Over the last half century, tree line advancements into tundra have also been documented in Russia (Esper and Schweingruber 2004), Canada (Lavoie and Payette 1994), and Alaska (Lloyd et al. 2003). However, fire and human activities in Russia have moved the tree line to the south in some areas (Vlassova 2002). In Russia, evergreen conifers have also been documented as expanding into the deciduous-conifer (larch) dominated zones of Siberia (Kharuk et al. 2005*b*). In mountainous areas of Scandinavia, the elevational tree line moved upslope during the last half of the 20th century in association with increases in temperature (Juntunen et al. 2002). Chapin et al. (2005) estimate that changes in shrub and forest expansion in the Arctic in recent decades have had minimal effect on the summer atmospheric (sensible and latent) heating associated with changes in albedo. This flux, $\sim$0.2 W$\cdot$m$^{-2}\cdot$decade^{-1}, is small compared to the energy exchange shifts of tundra that are associated with the albedo changes of a longer growing season. In this respect, there has not yet been a demonstrable feedback from shrub/forest expansion to climate. However, as will be discussed in *Future trajectories: Albedo and energy partitioning*, there are indications that such feedbacks could occur with the conversion of tundra to forest or even with a factor-of-two change in vegetative coverage and/or leaf area (Chapin et al. 2005, Zhang and Walsh 2007).

Increases in fire frequency have the potential to affect albedo and energy partitioning in the Arctic as they can shift landscapes dominated by low-albedo conifers toward a higher fraction of higher-albedo deciduous forests with lower carbon storage (Barrett et al. 2011). The area burned annually in the North America boreal region doubled from the 1960s–1970s to the 1980s–1990s (Kasischke and Turetsky 2006). In Alaska, large fire years have become more frequent in recent decades (Kasischke et al. 2010). Analyses of fire frequency data from Russia suggest a long-term average of annual area burned of about 10 million ha/yr (Wirth 2004), and subregional analysis of fire frequency data indicates that fires are becoming more frequent in Central Siberia (Kharuk et al. 2005*a*). Satellite-based analyses suggest that the area burned in Russia exceeded 10 million ha/yr from 1998 through 2003, with a peak of 22 million ha in 2003. Randerson et al. (2006) estimated that fires cause a decrease in energy exchange when averaged over an 80-year fire cycle (-2.3 ± 2.2 W/m^2), and Barrett et al. (2011) estimated that the most recent decade of fires in interior Alaska may have reduced the area occupied by black spruce by 4% and increased the area occupied by deciduous forest by 20%. The analysis of Euskirchen et al. (2009*b*) has suggested that this degree of change in vegetation associated with fire could cause a decrease in energy exchange of approximately 1 W/m^2 per decade over the region experiencing increased fire disturbance, resulting in a local climate feedback if the increase in fire occurrence was indeed driven by climatic warming.

Exchange of radiatively active gases with the atmosphere.—The exchange of radiatively active gases such as CO_2 and CH_4 between Arctic land ecosystems and the atmosphere is another important pathway through which these ecosystems could influence climate warming or cooling at the global scale. Northern high-latitude regions play an important role in the contemporary global dynamics of both CO_2 and CH_4. Analyses of atmospheric inversion models indicate that the region has been a sink for atmospheric CO_2 at up to 0.8 Pg C/yr in recent decades (Fig. 7) (McGuire et al. 2009), which is up to 25% of the net land/ocean flux of 3.2 Pg C/yr estimated for the 1990s by the IPCC Fourth Assessment Report (AR4) (Denman et al. 2007). Further inventory-based analyses have indicated that the land sink of this region has been between 0.3 and 0.6 Pg C/yr (McGuire et al. 2009), which is 30–60% of the 1.0 Pg C/yr global net land sink estimate for the 1990s. And the results of eddy covariance studies have indicated that terrestrial ecosystems in Alaska are acting as sinks for atmospheric CO_2 during the summer (Ueyama et al. 2013). An assessment of the carbon balance between 1990 and 2009 among observations, process models, and atmospheric inversion models has estimated that Arctic tundra has been an atmospheric CO_2 sink of 0.1 Pg C/yr, with an uncertainty between a 0.3 Pg C/yr sink and a 0.1 Pg C/yr source (McGuire et al. 2012). In boreal forest regions, Hayes et al. (2011) suggests that the strength of the northern high-latitude sink may be decreasing because of increased fire disturbance.

Carbon from high-latitude terrestrial ecosystems may be exported to freshwater and marine ecosystems through the loss of particulate and dissolved carbon, and this carbon may ultimately be released to the atmosphere. The Arctic Ocean drainage basin accounts for 11% of the global river water discharge from land to ocean (Lammers et al. 2001). Approximately 80 Tg C/yr is transferred from land to ocean via rivers in the northern cryosphere (Fig. 7), which is approximately 10% of the estimated 0.8 Pg C/yr transferred globally by rivers (Sarmiento and Gruber 2006). Coastal and wind erosion contribute another 8 Tg C/yr to the Arctic Ocean (McGuire et al. 2009). Kicklighter et al. (this issue) have estimated that dissolved organic carbon loading to river networks of the Arctic Basin has increased during the 20th century primarily because of climate-induced increases in runoff. It is important to understand the fate of this carbon as it becomes processed by freshwater aquatic ecosystems and is delivered to coastal ocean ecosystems.

Atmospheric inversion analyses have indicated that northern high latitude regions are a source of CH_4 to the atmosphere of between 15 and 50 Tg CH_4/yr (Fig. 7) (McGuire et al. 2009), which is between 3% and 9% of the net land/ocean source of 552 Tg CH_4/yr estimated by AR4 (Denman et al. 2007). In comparison with atmospheric inversion results, analyses based on observations have had higher lower and upper uncertainty bounds (32 and 112 Tg CH_4/yr, respectively) for the net source of CH_4 from the surface to the atmosphere in the northern cryosphere (McGuire et al. 2009). An assessment of the carbon balance of Arctic tundra between 1990 and 2009 among observations and process models has estimated that Arctic tundra has been a source of CH_4 to the atmosphere at 19 Tg C/yr, with an uncertainty between sources of 8 and 29 Tg C /yr (McGuire et al. 2012).

Methane emissions from the wetlands of northern terrestrial ecosystems are estimated to have increased by 0.6 Tg CH_4/yr between 1997 and 2006, primarily because of increased methane production in response to increases in precipitation in boreal Asia, and secondarily because of increases in temperature (McGuire et al. 2010). Although there has been much concern about how the expansion of thaw lakes in Siberia will influence global methane budgets (Walter et al. 2006), a recent study has suggested that the rate of expansion of thaw lakes is much less than estimated in previous studies and that current thaw lake expansion is having minimal effects on regional and global methane budgets (van Huissteden et al. 2011). In this respect, changes in methane release from the terrestrial Arctic represent a pathway of change (from the terrestrial biosphere to the atmosphere) though not yet a discernible feedback. The same statement applies to changes in CO_2 fluxes at the Arctic terrestrial surface.

Future trajectories

Albedo and energy partitioning.—Chapin et al. (2005) estimated that complete conversion of tundra to shrubs or to forest would potentially increase atmospheric summer warming in the Arctic more than would energy feedbacks associated with changes in the growing season. However, the importance of shrub and forest expansion in atmospheric warming in the future depends on their rates of expansion. Euskirchen et al. (2009*a*) applied a dynamic vegetation model of Arctic tundra in northern Alaska to estimate the time-dependent effects

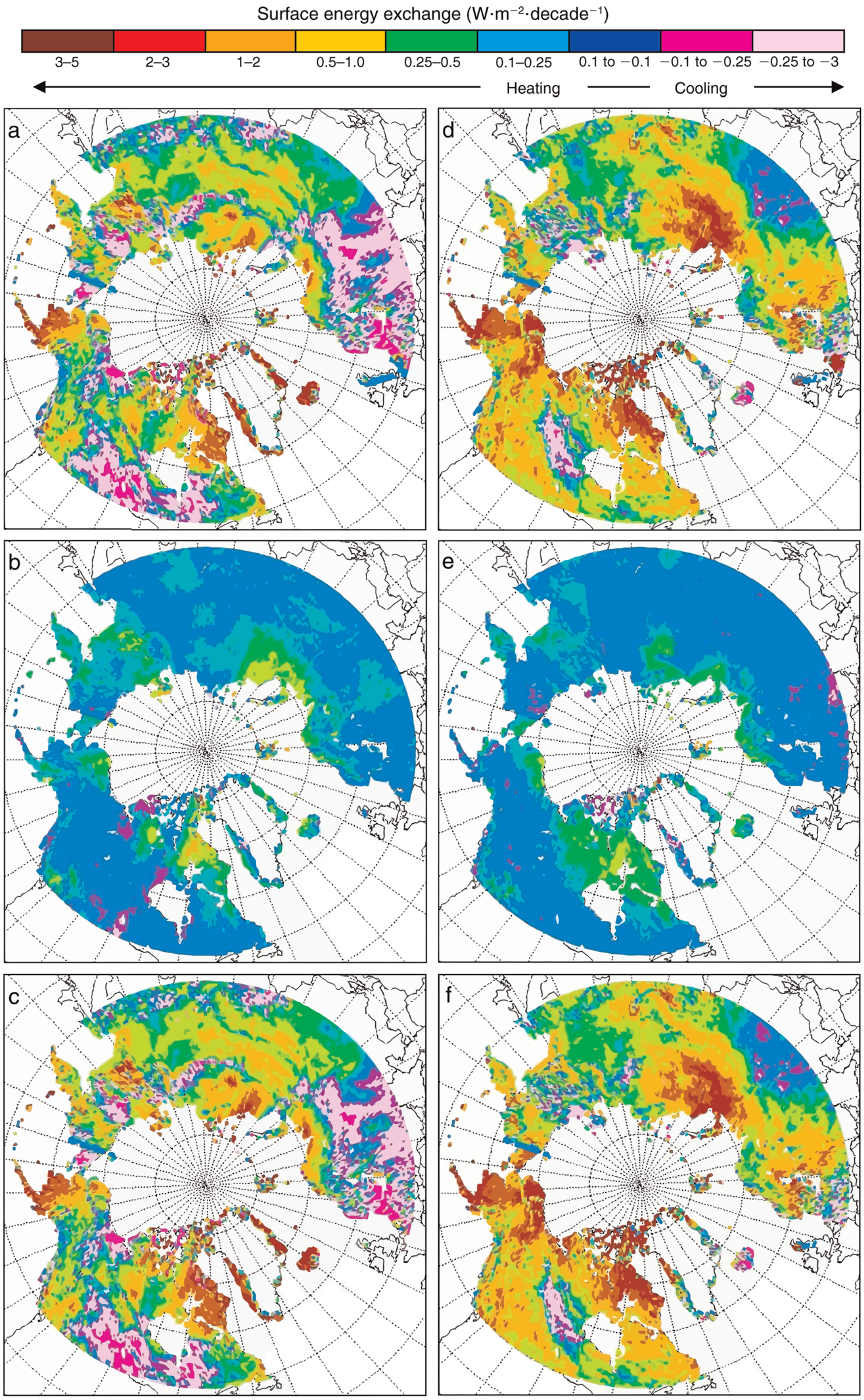

FIG. 6. Changes in surface energy exchange (i.e., changes in sensible and latent heat), due to changes in the timing of snowmelt for (a) 1910–1940 and (d) 1970–2000; changes in surface energy exchange due to changes in the timing of snow return (first snowfall in autumn) for (b) 1910–1940 and (e) 1970–2000; changes in surface energy exchange due to changes in the length of snow cover duration for (c) 1910–1940 and (f) 1970–2000. The figure is reprinted with permission from Euskirchen et al. (2007).

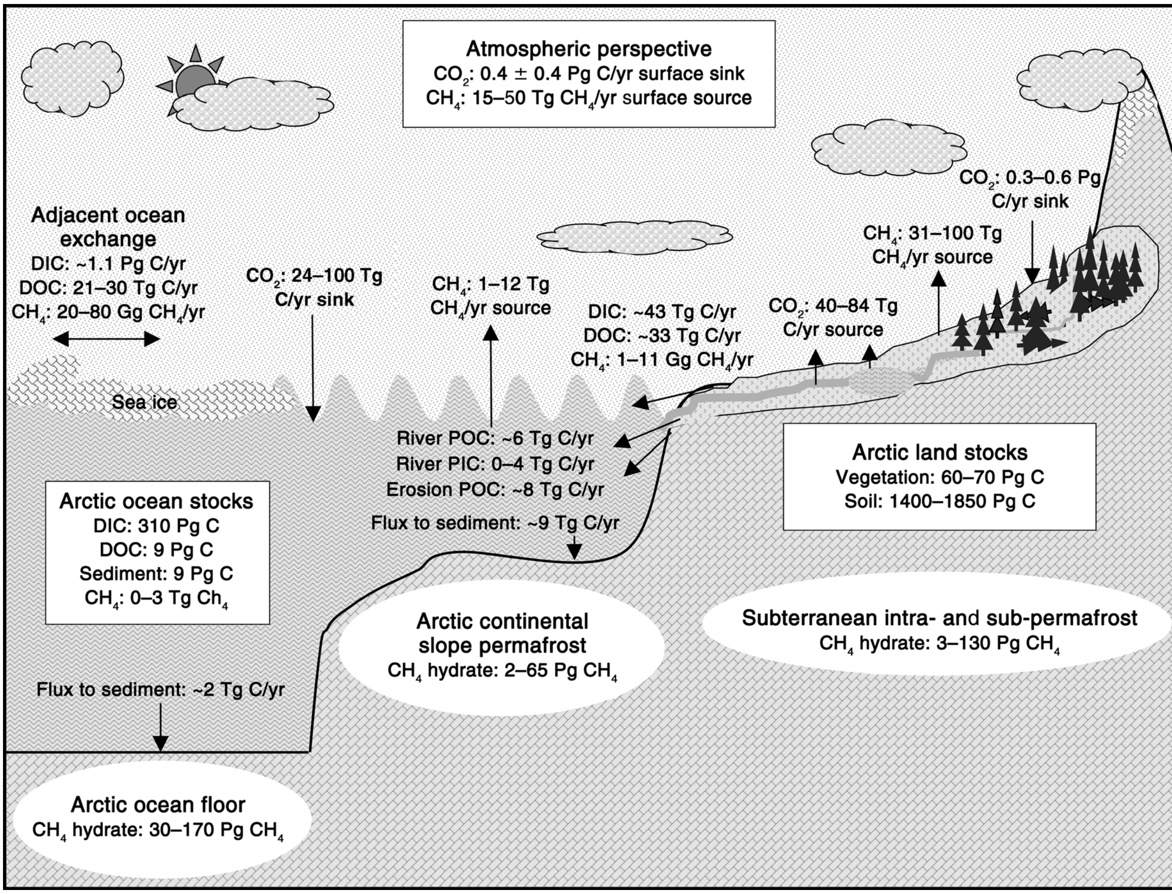

FIG. 7. The contemporary state of the carbon budget of the northern cyrosphere region, presented by McGuire et al. (2009). Abbreviations are: DIC, dissolved inorganic carbon; DOC, dissolved organic carbon; PIC, particulate inorganic carbon; POC, particulate organic carbon.

of the expansion of shrubs on atmospheric heating budgets in response to projected climate change for the region. In these simulations, plant production increased to cause greater aboveground biomass, which decreased summer albedo and resulted in greater regional heat absorption (0.3 ± 0.2 $W \cdot m^{-2} \cdot decade^{-1}$). However, the decrease in albedo due to a shorter snow season (5.1 ± 1.6 d/decade) resulted in much greater regional heat absorption (3.3 ± 1.2 $W \cdot m^{-2} \cdot decade^{-1}$) than that associated with increases in vegetation carbon. The rate of forest expansion (tree line advancement) is very slow (Lloyd et al. 2003), and the effects of these changes on atmospheric warming will likely take several hundred to 1000 years or more to become significant. Thus, the decrease in snow cover duration is likely to be the primary factor responsible for decreases in terrestrial surface albedo (Table 1).

Future changes in climate are likely to have pronounced effects on fire regime in northern high latitudes (Flannigan et al. 2000, 2005). Fire frequency in Alaska and Canada could increase four to six times depending on emissions scenarios (Balshi et al. 2009). Euskirchen et al. (2009*b*) have estimated that changes in summer heating due to changes in the forest stand age distributions under future fire regimes in the western Arctic would reduce energy exchange because of increases in summer albedo (–0.9 $W \cdot m^{-2} \cdot decade^{-1}$ across climate scenarios; Table 1). This was substantially less, however, than the 4.3 $W \cdot m^{-2} \cdot decade^{-1}$ atmospheric heating due to changes in the length of the growing season.

Exchange of radiatively active gases with the atmosphere.—Only a few modeling studies have examined how large-scale ecosystem carbon storage might be affected by exposing carbon in thawing permafrost to decomposition, an issue that has been referred to as permafrost carbon feedback (Zhuang et al. 2006, Zimov et al. 2006*a*, *b*). Among the studies that have explicitly considered the exposure of carbon to decomposition, the change in soil carbon from the northern cryosphere region ranges between an increase of 14 Pg C to a release of 114 Pg C by 2100 (Zhuang et al. 2006, Koven et al. 2011, Schaefer et al. 2011, Schneider von Deimling et al. 2012). While there are some differences among the models with respect to the magnitude of carbon loss in response to climate change, the models are consistent in that they all estimate that there is a qualitatively different response when the effects of thaw on the decomposition of carbon stored in permafrost is considered (Table 1). The coupled model simulations of Schneider von Deimling et al. (2012) have estimated that the loss of carbon from permafrost soils leads to additional warming of between 0.04° and 0.23°C by 2100. While technically representing a feedback from the atmosphere to the terrestrial biosphere, such temperature changes are much smaller than the anticipated direct anthropogenic warming of several °C in the Arctic by 2100.

It is important to recognize that the range of carbon release estimated by models has been much less than the 232–380 Pg C range of release by 2100 estimated from a recent survey of experts on permafrost thaw and carbon release (Schuur et al. 2011). The modeling analyses consider only vertical thaw of permafrost and do not consider losses of carbon that might be associated with the thaw and subsidence of ice-rich permafrost (i.e., thermokarst collapse). The representation of thermokarst development and its effects on the soil carbon storage of northern high-latitude regions remains a substantial challenge for the large-scale modeling community. Several modeling analyses have indicated that climate change has the potential to substantially increase biogenic CH_4 emissions throughout the northern cryosphere during the 21st century (Zhuang et al. 2006, Koven et al. 2011, Schneider von Deimling et al. 2012; Table 1). However, the effects of thermokarst on biogenic CH_4 emissions have not been adequately considered in these models, and the release of CH_4 could be greater than projected. It is estimated that thaw-lake expansion could increase CH_4 emissions from thaw lakes by 50%, but that this would be responsible for only about 4% of total CH_4 emissions from northern wetlands (van Huissteden et al. 2011). Also, the effects of changes in wetland area in response to permafrost thaw have the potential to affect methane emissions. The signal is likely to be quite complicated because of the higher methane emissions associated with initial wetland expansion as permafrost thaws, followed by lower emissions as thawed areas drain (Avis et al. 2011). However, the long-term effect could be a reduction in methane emissions because of reduced wetland area. If the latter scenario were to occur, a negative feedback associated with CH_4 emissions to the climate system would follow.

Feedback Pathways of Arctic Marine Ecosystems

Observed changes

Arctic Ocean ecosystems may also influence the climate system by affecting the atmosphere-ocean-sea ice exchange of energy and greenhouse gases. At the same time, Arctic marine biogeochemistry and ecology are sensitive to global climate change. Exchanges of energy involve both complex atmospheric chemistry-aerosol-cloud interactions linked to ocean emissions of aerosol precursor gases (Gabric et al. 2005) and particles (Leck and Bigg 2005), and the absorption, potentially, of solar heat by phytoplankton (Lengaigne et al. 2009) and sea-ice microalgae. Many of these processes, which are sensitive to climate change in the terrestrial regions of the Arctic, impact ocean biogeochemistry by modifying the delivery and transport pathways of freshwater and particulate and dissolved materials. Similarly, physical features of the atmosphere (e.g., wind regimes, warming), ocean (e.g., water mass pathways, eddy transport), and sea ice (i.e., distribution, thickness, melt dynamics) determine the variety of physical (e.g., light penetration), and chemical (e.g., nutrient supply) settings of the Arctic seas that support the polar marine food web and its interactions with biogeochemical cycles. The Arctic Ocean is an important sink for atmospheric CO_2, and a strong source of methane and other biogenic, climate-relevant gases, such as dimethylsulfide (DMS). How the spatial pattern of sinks and sources will change in the future is a crucial question. Even though Arctic observations in the marine biological domain are sparse and unevenly distributed in space and time, we are beginning to observe changes.

There is evidence that the productivity of the Arctic Ocean is changing. Changes in marine primary production impact food webs, biogeochemical cycling, and air–sea fluxes of biogenic gases. Field measurements have suggested that primary productivity in the Bering Strait and Chukchi Sea has declined in recent years compared to previous decades (Lee et al. 2012). However, considering temporal and geographic variations, it is not clear if this is a trend. Remote-sensing estimates of primary production have displayed an overall increase in the Amerasian Arctic and small decreases in the Canadian and European sectors for the year 2007, compared to the 1998–2002 mean (Arrigo et al. 2008). On the pan-Arctic scale, annual primary production increased 23% in 2007 compared to the 1998–2002 mean (Arrigo et al. 2008). Numerical modeling and remote sensing algorithms have revealed that more open-water area has resulted in a longer phytoplankton growing season, coinciding with increased Arctic Ocean primary production over the last decade (Arrigo et al. 2008, Jin et al. 2012). However, in relation to increased open-water area, numerical models illustrate the roles of nutrient limitation and enhanced light availability in open water (Popova et al. 2012). A recent analysis of published data has indicated that nitrogen supply is the dominant factor limiting primary production per unit area in seasonally ice-covered seas (Tremblay and Gagnon 2009). Disagreement among models is attributed to variations in the simulated vertical mixing that controls nutrient supply. If an increase in biotic uptake is sufficient to have a significant effect on ambient CO_2 concentrations, this would result in a negative feedback on climate. A study using several model-based tools to analyze the dynamics of the Arctic Basin concluded that the Arctic Ocean was a net CO_2 sink of 57.8 Tg C/yr between 1997 and 2006 (McGuire et al. 2010). This is in general agreement with the synthesis of observations conducted by Bates and Mathis (2009), which estimated Arctic Ocean C uptake ranging from 66 to 199 Tg C/yr and contributing 5–14% of the net global CO_2 flux. Recently, a high-resolution survey of CO_2 partial pressure across the Chukchi Sea and Canadian Basin conducted after the recent major sea-ice retreat (summers 2007–2009) has shown a "great increase" of sea surface CO_2 concentration relative to earlier observations (Cai et al. 2010). These findings, along with detailed analysis, suggest an equilibration of surface waters with the atmosphere upon the reduction of sea ice in the central Arctic basin.

There are numerous examples of recently observed changes in species distribution ranges, biomasses, and trophic cascades (e.g., Grebmeier et al. 2010, Bluhm et al. 2011). Most of the reported changes have been near the Arctic margins, instead of in the central Arctic, and they have described distribution shifts in marine mammals and declines in seabird populations. Distribution shifts at lower trophic levels have also occurred and have often been tied to changes in water temperature and/or water mass transport. For example, under warming conditions, the Atlantic phytoplankton *Emiliania huxleyi* has moved into the high Arctic from the southern Barents Sea (Hegseth and Sundfjord 2008). Community-wide northward shifts of fish and invertebrate taxa in the Bering Sea have also been attributed to warming water (Mueter and Litzow 2008).

Plankton species shifts due to decadal climate variations and sea-ice loss have been observed in the Bering Sea (e.g., Coyle et al. 2008) and revealed in model simulations (Jin et al. 2009). Earlier sea-ice loss in the Bering Sea has altered the tight link between the water column and the seafloor biological realms (pelagic-benthic coupling) that are characteristic of seasonal ice zones (e.g., Hunt et al. 2002). This loss has shifted the benthos-oriented ecosystem to a phytoplankton-zooplankton-dominated system in the Northern Bering Sea (Grebmeier et al. 2006). The resulting reduction in benthic prey populations, along with reduced sea ice, has already displaced zooplankton, fish, and marine mammal populations northward (Grebmeier et al. 2010). However, the past few years (2008–2012) have also seen a return to colder ocean temperatures and increased winter sea ice cover in the Bering Sea.

There is also observational evidence for a shift to smaller algal species in Canadian Basin areas once covered by perennial pack ice due to freshening, and consequently, changes limiting nutrient supply (Li et al. 2009). Because the fraction of particulate organic carbon (POC) to dissolved organic carbon (DOC) produced is different for large-sized algae, a change to smaller cells shifts this POC/DOC ratio, impacting carbon export, pelagic-benthic coupling, and benthic populations. Large-scale numerical modeling results corroborate the observed shift in size and furthermore display non-uniform primary production responses across the Arctic, due to diminishing sea ice (Jin et al. 2012).

Ocean acidification further complicates the effects of warming and reduced sea ice on marine ecosystems. Cold temperatures and ocean mixing patterns cause the Arctic Ocean to have naturally low pH and calcium carbonate ($CaCO_3$) mineral saturation states, relative to temperate and tropical regions (Fabry et al. 2009). Freshwater contributions (from sea ice melt and river runoff) and high rates of primary production on Arctic shelves (in combination with increased atmospheric CO_2) are driving forces contributing to undersaturation of $CaCO_3$ (Table 1), and this undersaturation is detrimental for carbonate-forming phytoplankton, zooplankton, and benthic species. $CaCO_3$ (i.e., aragonite) suppression has already been observed in the Subarctic Pacific Ocean (Mathis et al. 2011), the Canadian Basin, the Canadian Archipelago and Beaufort Sea shelf, and some subsurface waters of the Chukchi Sea (Bates and Mathis 2009).

In regard to the potent greenhouses gas methane, there have been recent observations of methane leakage through shallow East Siberian Arctic Shelf sediments due to perforated subsea permafrost. Shakhova et al. (2010) have estimated the recent release to be 7 Tg/yr, which is at the same magnitude as estimates of methane venting from the entire remainder of the world ocean.

Future trajectories

An evaluation of the future trajectory of marine ecosystems in the Arctic requires improved understanding of the linkages and feedbacks between physical drivers, biological production, and biogeochemical cycles in the Arctic Ocean. The key processes and responses to change will likely vary within different Arctic sub-regions (Carmack et al. 2006). Such differences may be attributed to the spatial heterogeneity of the physical environment (e.g., bathymetry, proximity to land), climate forcing (e.g., wind regimes, sea ice loss), and ecosystems.

Changes in ocean gateway inflows, Arctic watersheds, and land margins may significantly influence the carbon cycle of the nearly land-locked Arctic Ocean, with over half its area underlain by shallow continental shelves. Even a small change in the dominant Arctic Ocean carbon exchange with adjoining oceans ($\sim$3 Pg C/yr) could have significant and rapid impact on the balance of Arctic Ocean CO_2 sinks and sources (Bates and Mathis 2009). Northern rivers deliver a substantial amount of carbon to the Arctic Ocean and its marginal seas: net land to ocean organic and inorganic carbon flux is $\sim$76 Tg (Fig. 7). Increased riverine discharge and coastal erosion associated with reduced sea ice cover and coastal permafrost degradation will add to this amount. Much of the dissolved organic carbon will likely decompose rapidly in coastal waters, while a significant fraction of the particulate matter may add to the current $\sim$9 Tg C/yr sediment flux to the Arctic Continental

Slope. In heavily impacted, shallow interior shelf seas, greater export of chromophoric dissolved organic matter and sediment-laden waters from land and coastal regions would attenuate light even more strongly, limiting photosynthesis (Retamal et al. 2008, Semiletov et al. 2011). The terreginous sediment that sinks to the benthos becomes an ocean sediment source of CO_2. Under warming ocean temperatures, methane from subsea permafrost and methane hydrates also have the potential to change the carbon fluxes from Arctic Ocean sediments, with a positive climate feedback.

Changes in Arctic Ocean warming and circulation, defined in the broadest sense, also impact Arctic Ocean ecosystems and biogeochemical cycles. Increasing temperatures directly affect biological rates (e.g., microbial respiration, zooplankton grazing). The distribution of properties or processes of importance to biological productivity, such as nutrient supply, light penetration, and cycling and remineralization of organic matter, can be altered by changes in circulation, mixing, and freshwater balance, as well as the thickness and distribution of sea ice and snow cover. Key processes affecting these distributions include shelf-basin transport, vertical mixing and diffusion, tidal mixing, and freshwater input.

Reduced sea ice cover and higher temperatures also facilitate ocean–air–sea ice interactions. Loss of sea ice exposes undersaturated surface waters to the atmosphere, thus potentially increasing the uptake of CO_2. Furthermore, enhanced opening of polynyas (areas of open water surrounded by sea ice) and leads may impact inorganic carbon flux through brine-rejection during deep-water formation. As a result of increased melting of sea ice cover and river discharge, there will likely be changes in air-sea, sea-ice, polynya/lead, sea-ice melt, and river-water gas exchanges. Warming temperatures have also been found to promote the transport of biogenic gases (e.g., CO_2, DMS) through sea ice (Delille et al. 2007). The cumulative effect of these exchange processes is highly uncertain. More studies are needed to quantify these uncertain and opposing processes and pathways, and to determine whether ongoing changes are actually enhancing feedbacks.

There is still considerable uncertainty as to whether the Arctic Ocean will become more productive, and a larger net sink of CO_2 in the future (Table 1). Three of the four global coupled carbon-cycle–climate models analyzed by Steinacher et al. (2010) project an increase in arctic marine net primary productivity over the 21st century. It appears likely that on short timescales, increased uptake of CO_2 will continue, due to further sea-ice loss and high rates of primary production (Bates and Mathis 2009). To the contrary, there have been recent observations suggesting the potential for a reversal of this trend, as some Canadian Basin waters have become saturated with CO_2 (Cai et al. 2010). Based on detailed observational analyses, Cai et al. (2010) concluded that the Arctic Ocean will not become a larger atmospheric CO_2 sink under ice-free conditions. Such a transition would possibly have long-term consequences for the atmospheric CO_2 sink, through decreasing carbon burial and sequestration on Arctic continental shelves and deep basins.

With changes in CO_2 concentrations, sea-ice distribution and thickness, temperatures, snow cover, freshwater releases, and other environmental conditions will come changes in Arctic marine ecosystems (ACIA 2005). Although global and regional model simulations of contemporary pan-Arctic ocean primary production show features similar to observations, their predictive capabilities diminish with the retreat of sea ice into the future (Popova et al. 2012), and they do not capture the complex dynamics of Arctic marine ecosystems. Loss of habitats will also result from retreating sea ice (e.g., Arrigo et al. 2008). However, model results using the Los Alamos community ice model (CICE) and including Arctic ice algal production (Deal et al. 2011) suggest a near term ($<$10 years) overall increase in annual ice algal primary production, due to the trend toward thinner sea ice cover (Jin et al. 2012). Rather than simply being proportional to ice area alone, the combined effects of ice growth/melt, nutrient availability, and light intensity determined the simulated annual sea-ice algal production within the Arctic Circle from 1998 to 2007.

Further, Gabric et al. (2005) analyzed the impact of reduced sea-ice cover on the simulated change of DMS flux and the consequent future aerosol production and radiative budget of the Arctic. Their results suggested a substantial DMS-derived cooling effect over the Arctic Ocean in a scenario of atmospheric CO_2 tripling by 2080 (Gabric et al. 2005). This change represents a potential negative feedback to climate warming and sea-ice loss due to higher DMS production, and thus increased sulfate aerosols and cloud condensation nuclei. Other sources of these nuclei (and organic aerosols) produced by biological activities (i.e., organic particles, dissolved organic matter, and volatile organic compounds) may influence cloud condensation nuclei properties and similarly feed back to climate (Leck and Bigg 2005). Changes in Arctic Ocean emissions of biogenic aerosol precursors, such as DMS and other organics, are arguably the most immediate possibility for an atmospheric feedback involving Arctic marine ecosystems.

In addition to the physical and biotic responses briefly discussed above, several mediating processes may exacerbate the sensitivity of certain species to climate change (e.g., see Bates and Mathis 2009; Box 3 in Deal et al., *in press*). These disturbances include increased use of resources (e.g., fisheries, resource exploration and development) and ocean acidification as consequences of the uptake of anthropogenic CO_2. High-latitude seas are considered a "bellwether" for the impacts of ocean acidification on marine organisms (Fabry et al. 2009). With current CO_2 emission rates, models project expanded areas of aragonite undersaturation, including some Arctic surface waters during the next several decades (Steinacher et al. 2009, Denman et al. 2011). Benthic ecosystems are expected to be most negatively impacted by ocean acidification, with potentially profound implications for Arctic marine ecosystems (Bates and Mathis 2009).

Conclusions

In the last 50 years, a wide range of changes in the Arctic have been documented. The Arctic atmospheric warming trend in the last decade was exceptionally strong, reaching 1.35°C/decade, ~1.6 times stronger than the warming rate of the Northern Hemisphere. The warming of intermediate layers of the Arctic Ocean in particular was exceptional, with no analog in the history of instrumental observations. Changes in atmospheric circulation have been important drivers of the regional pattern of warming within the Arctic, especially over the northern land areas. Atmospheric circulation also accounts for about one-third of the downward trend in the summer ice-area minima and an even greater fraction of the year-to-year variability. Polar amplification has been shown to arise from both local contribution (involving the ice–albedo–temperature feedback) and remote contribution (achieved through the poleward transport of energy and moisture from lower latitudes). The Canadian Basin has transitioned from an ice-production region to an ice-melting one, a remarkable change of status in response to global warming. In addition, unlike other areas in the Arctic Ocean, it appears the waters of the Chukchi Sea and Canadian Basin have become saturated with CO_2. Emerging signs of the response of high-latitude regions to Arctic ice loss are apparent in Arctic surface air temperatures, especially in autumn; in tundra vegetation; and possibly in early-winter atmospheric circulation anomalies that extend into middle latitudes. The most demonstrable feedback is the warming of the lower atmosphere over the maritime Arctic in autumn/winter and over the terrestrial Arctic in spring. Both are at least partially manifestations of the ice/snow–albedo–temperature feedback, resulting in this case from reductions in sea ice and terrestrial snow cover.

Numerous and generally consistent changes have also been documented in the permafrost, glacier and ice sheet, and hydrologic systems. Warming has been detected in permafrost throughout the Circumpolar North, with few exceptions. To a lesser extent, increases in active layer thickness have also been reported. Observed hydrological responses to permafrost degradation include shrinking lakes in discontinuous permafrost regions and an increased number of small ponds in continuous permafrost regions. Other hydrologic trends include increased river runoff in Eurasia and North America, earlier snowmelt, and decreased snow cover area. Published literature has reported increasing and decreasing trends in precipitation, though analyses of evapotranspiration generally point to increases. The Greenland ice sheet melt extent has increased since 1972, leading to both enhanced surface-ice loss and enhanced total ice-mass loss from Greenland of 430 Gt/yr for 2010–2011, with potentially important consequences for ocean density changes and the thermohaline circulation, and specifically for the Atlantic Meridional Overturning Circulation and its impact on the climate system. Also, mass balance observations on glaciers and ice caps for recent decades have shown an overall increase in mass loss, trending toward greater negative annual surface mass balance during the first half (2001–2005) of the first decade of the 21st century, while few glaciers and ice caps have increased in mass.

Current understanding of the carbon cycle in terrestrial regions of the Arctic is limited by considerable uncertainties. Integrated studies of regional carbon dynamics are needed to provide better information on key elements of the carbon cycle in this region. Such studies should link observations of carbon dynamics to the processes that influence these dynamics. Furthermore, the rapidity and extent of change occurring in terrestrial regions of the Arctic demand increased attention to collection of appropriate time-series data for the carbon cycle of this region. The resulting information should be incorporated into modeling efforts that connect carbon dynamics and climate. These studies should focus on sensitive parts of the system, for example, areas experiencing major changes or crossing thresholds, such as increased fire disturbance, permafrost degradation, and changes in hydrology.

One of the prevailing debates among Arctic marine scientists centers on whether climate forcing will increase or decrease marine productivity and whether the Arctic Ocean will remain a net CO_2 sink into the future. There is clear evidence that marine primary productivity is changing and that habitats and trophic structures are shifting, coincident with sea-ice loss. This is not surprising, as in the Arctic Ocean sea-ice thickness, extent, and melt dynamics determine light availability and nutrient supply, which then affect primary productivity and biogeochemical cycling. Diminishing sea ice has the potential to shift the marine ecosystem to a new state. Much more baseline and background data are critically needed to drive models and investigate interactions. Because of concurrent changes in multiple parameters and synergisms with other human perturbations (e.g., fisheries), we must address these issues together and assess the mechanisms behind the responses. As with terrestrial carbon dynamics, marine studies should focus on sensitive parts of the system, for example, those such as ocean acidification that appear to be especially close to "tipping" points. Furthermore, the challenge of scaling up inherently fine-scale biological and chemical observations to address system-scale questions must be tackled through the concerted efforts of observationalists and modelers.

Of the feedbacks considered here, evidence in ongoing changes is most compelling for surface albedo–temperature feedback, which is amplifying temperature changes over land (primarily in spring) and ocean (primarily in autumn–winter). Fluxes of methane released from the marine and terrestrial Arctic represent pathways of change, though not yet a discernible feedback. Other feedbacks likely to emerge are those involving surface fluxes of trace gases, changes in the distribution of vegetation, changes in atmospheric water vapor arising from higher temperatures and greater areas of open ocean, impacts of Arctic freshwater fluxes upon the meridional overturning circulation of the ocean, and changes in Arctic clouds arising from increases in dimethyl sulfide and marine biogenic aerosols. To varying degrees, each of these feedbacks involve Arctic terrestrial and marine ecosystems, which will be affected in important ways by the ongoing and future changes.

Regardless of the driving forces, the combination of recent observations and historical information suggests that the Arctic system may be entering a state not seen in the historical record. The complex interplay of physical, chemical, biological, and social processes interact to such a degree that it is not possible to understand future trajectories without developing more fully holistic perspectives of the complete system, including its feedbacks. Moreover, there will almost certainly be surprises in the emergence of new and key feedbacks, some of which may not have been addressed in this review. Whether or not these feedbacks are anticipated, their detection and understanding may be regarded as grand challenges of Arctic system science. The optimal design of observational networks and the enhanced sophistication and credibility of models will be required to meet these challenges.

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Literature Cited

ACIA. 2005. Arctic climate impact assessment. Cambridge University Press, Cambridge, UK.

Åkerman, H. J., and M. Johansson. 2008. Thawing permafrost and thicker active layers in sub-Arctic Sweden. Permafrost and Periglacial Processes 19:279–292.

Arrigo, K. R., G. van Dijken, and S. Pabi. 2008. Impact of a shrinking Arctic ice cover on marine primary production. Geophysical Research Letters 35:L19603.

Avis, C. A., A. J. Weaver, and K. J. Meissner. 2011. Reduction in areal extent of high-latitude wetlands in response to permafrost thaw. Nature Geoscience 4:444–448.

Balshi, M. S., A. D. McGuire, P. Duffy, M. Flannigan, J. Walsh, and J. Melillo. 2009. Assessing the response of area burned in western boreal North America using a multivariate adaptive regression splines (MARS) approach. Global Change Biology 15:578–600.

Barrett, K., A. D. McGuire, E. E. Hoy, and E. S. Kasischke. 2011. Potential shifts in dominant forest cover in interior Alaska driven by variations in fire severity. Ecological Applications 21:2380–2396.

Bates, N. R., and J. T. Mathis. 2009. The Arctic Ocean marine carbon cycle: evaluation of air-sea CO_2 exchanges, ocean acidification impacts and potential feedbacks. Biogeosciences 6:6695–6747.

Bekryaev, R. V., I. V. Polyakov, and V. A. Alexeev. 2010. Role of polar amplification in long-term surface air temperature variations and modern Arctic warming. Journal of Climate 23(14):3888–3906.

Belchansky, G. I., D. C. Douglas, and N. G. Platonov. 2008. Fluctuating Arctic sea ice thickness changes estimated by an in situ learned and empirically forced neural network model. Journal of Climate 21:716–729.

Bhatt, U. S., et al. 2010. Circumpolar Arctic tundra vegetation change is linked to sea-ice decline. Earth Interactions 14:1–20.

Bluhm, B. A., A. V. Gebruk, R. Gradinger, R. R. Hopcroft, F. Huettmann, K. N. Kosobokova, B. I. Sirenko, and J. M. Weslawski. 2011. Arctic marine biodiversity: An update of species richness and examples of biodiversity change. Oceanography 24(3):232–248.

Bolton, W. R., L. D. Hinzman, and K. Yoshikawa. 2004. Water balance dynamics of three small catchments in a Sub-Arctic boreal forest. Pages 213–223 *in* D. L. Kane and D. Yang, editors. Northern research basins water balance. Proceedings of a workshop held at Victoria, Canada, March 2004. IAHS Series of Proceedings and Reports Number 290. IAHS, Wallingford, UK.

Box, J. E., and D. T. Decker. 2011. Greenland marine-terminating glacier area changes: 2000–2010. Annals of Glaciology 52(59):91–98.

Brown, J., O. J. Ferrians, Jr., J. A. Heginbottom, and E. S. Melnikov, editors. 1997. Circum-Arctic map of permafrost and ground-ice conditions. Circum-Pacific Map Series CP-45, scale 1:10,000,000, 1 sheet. U.S. Geological Survey in Cooperation with the Circum-Pacific Council for Energy and Mineral Resources, Washington, D.C., USA.

Brown, R., C. Derksen, and L. Wang. 2010. A multi-data set analysis of variability and change in Arctic spring snow cover extent, 1967–2008. Journal of Geophysical Research 115:D16111.

Brown, R. D., P. Y. Groisman, N. Labba, and V. Radionov. 2012. Changing snow cover and its impacts. Chapter 4 *in* Arctic Monitoring and Assessment Programme (AMAP), editors. Snow, water, ice and permafrost in the Arctic (SWIPA): Climate change and the cryosphere. AMAP, Oslo, Norway.

Brown, R. J. E., and T. L. Péwé. 1973. Distribution of permafrost in North America and its relationship to the environment, a review, 1963–1973. Pages 71–100 *in* National Academy of Sciences, editors. Proceedings of the Second International Conference on Permafrost. National Academy of Sciences, Washington D.C., USA.

Bryden, H. L., H. R. Longworth, and S. A. Cunningham. 2005. Slowing of the Atlantic meridional overturning circulation at 25°N. Nature 438:655–657.

Cai, W., et al. 2010. Decrease in the CO_2 uptake capacity in an ice-free Arctic Ocean Basin. Science 329:556–559.

Cappelen, J. 2011. DMI monthly climate data collection 1768–2010, Denmark, the Faroe Islands and Greenland. Technical Report 11-05. Danish Meteorological Institute, Copenhagen, Denmark.

Carmack, E., D. Barber, J. Christensen, R. Macdonald, B. Rudels, and E. Sakshaug. 2006. Climate variability and physical forcing of the food webs and the carbon budget on panarctic shelves. Progress in Oceanography 71:145–181.

Cazenave, A., and W. Llovel. 2010. Contemporary sea level rise. Annual Review of Marine Science 2:145–173.

Chapin, F. S., III, et al. 2005. Role of land-surface changes in Arctic summer warming. Science 310:657–660.

Christiansen, H. H., and O. Humlum. 2008. Interannual variations in active layer thickness in Svalbard. Pages 257–262 *in* D. L. Kane and K. M. Hinkel, editors. Ninth International Conference on Permafrost Proceedings. Institute of Northern Engineering, University of Alaska, Fairbanks, USA.

Cogley, G. 2012. The future of the world's glaciers. Pages 197–222 *in* A. Henderson-Sellers and K. McGuffie, editors. The future of the world's climate. Elsevier, Amsterdam, The Netherlands.

Comiso, J. C. 2003. Warming trends in the Arctic from clear sky satellite observations. Journal of Climate 16:3498–3510.

Coyle, K. O., A. I. Pinchuk, L. B. Eisner, and J. M. Napp. 2008. Zoo-plankton species composition, abundance and biomass on the eastern Bering Sea shelf during summer: the potential role of water column stability and nutrients in structuring the zooplankton community. Deep-Sea Research Part II: Topical Studies in Oceanography 55:1775–1791.

Deal, C., M. Jin, S. Elliott, E. Hunke, M. Maltrud, and N. Jeffery. 2011. Large-scale modeling of primary production and ice algal biomass within Arctic sea ice in 1992, Journal of Geophysical Research 116:C07004.

Deal, C. J., et al. 2012. Progress and challenges in biogeochemical modeling of the Pacific Arctic sector. *In* J. Grebmeier and W. Maslowski, editors. The Pacific Arctic region: ecosystem status and trends in a rapidly changing environment. Springer Publishing, New York, New York, USA.

Delille, B., B. Jourdain, A. Borges, J. L. Tison, and D. Delille. 2007. Biogas (CO_2, O_2, dimethylsulfide) dynamics in spring Antarctic fast ice. Limnology and Oceanography 52:1367–1379.

Denman, K., J. Christian, N. Steiner, H. O. Portner, and Y. Nojiri. 2011. Potential impacts of future ocean acidification on marine ecosystems and fisheries. ICES Journal of Marine Science 68:1019–1029.

Denman, K. L., et al. 2007. Couplings between changes in the climate system and biogeochemistry. Pages 499–587 *in* S. Solomon, D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Avery, M. Tignor, and H. L. Miller, editors. Climate change 2007: the physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, UK.

Déry, S. J., and R. D. Brown. 2007. Recent Northern Hemisphere snow cover extent trends and implications for the snow albedo feedback. Geophysical Research Letters 34:L22504.

Déry, S. J., M. A. Hernández-Henríquez, J. E. Burford, and E. F. Wood. 2009. Observational evidence of an intensifying hydrological cycle in northern Canada. Geophysical Research Letters 36:L13402.

Dickson, R. R., R. Curry, and I. Yashayaev. 2003. Recent changes in the North Atlantic. Philosophical Transactions of the Royal Meteorological Society A 361:1917–1934.

Dmitrenko, I. A., I. V. Polyakov, S. A. Kirillov, L. A. Timokhov, I. E. Frolov, V. T. Sokolov, H. L. Simmons, V. V. Ivanov, and D. Walsh. 2008. Toward a warmer Arctic Ocean: spreading of the early 21st century Atlantic Water warm anomaly along the Eurasian Basin margins. Journal of Geophysical Research 113:C05023.

Dyurgerov, M. B. 2010. Data of glaciological studies—reanalysis of glacier changes: from the IGY to the IPY, 1960–2008. Publication No. 108. Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado, USA.

Dyurgerov, M. B., and M. F. Meier. 2005. Glaciers and the changing Earth system: a 2004 snapshot. Occasional Paper No. 58. Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado, USA.

Esper, J., and F. H. Schweingruber. 2004. Large-scale treeline changes recorded in Siberia. Geophysical Research Letters 3:L06202.

Ettema, J., M. R. van den Broeke, E. van Meijgaard, W. J. van den Berg, J. L. Bamber, J. E. Box, and R. C. Bales. 2009. Higher surface mass balance of the Greenland ice sheet revealed by high-resolution climate modeling. Geophysical Research Letters 36:L12501.

Euskirchen, E. S., A. D. McGuire, and F. S. Chapin, III. 2007. Energy feedbacks of northern high-latitude ecosystems to the climate system due to reduced snow cover during 20th Century warming. Global Change Biology 13:2425–2438.

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

Euskirchen, E. S., 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, T. S. Rupp, F. S. Chapin, III, and J. E. Walsh. 2009*b*. Projected changes in atmospheric heating due to changes in fire disturbance and the snow season in the western Arctic, 2003–2100. Journal of Geophysical Research: Biogeosciences 114:G04022.

Fabry, V. J., J. B. McClintock, J. T. Mathis, and J. M. Grebmeier. 2009. Ocean acidification at high latitudes: the bellwether. Oceanography 22:160–171.

Fettweis, X., E. Hanna, H. Gallee, P. Huybrechts, and M. Erpicum. 2008. Estimation of the Greenland ice sheet surface mass balance for the 20th and 21st centuries. Cryosphere 2:117–129.

Flannigan, M. D., K. A. Logan, B. D. Amiro, W. R. Skinner, and B. J. Stocks. 2005. Future area burned in Canada. Climatic Change 72:1–16.

Flannigan, M. D., B. J. Stocks, and B. M. Wotton. 2000. Climate change and forest fires, Science of the Total Environment 262:221–229.

Forbes, D. L., editor. 2011. State of the Arctic coast 2010—Scientific review and outlook. International Arctic Science Committee, Land-Ocean Interactions in the Coastal Zone, Arctic Monitoring and Assessment Programme, International Permafrost Association. Helmholtz-Zentrum, Geesthacht, Germany.

Francis, J. A., W. Chan, D. J. Leathers, J. R. Miller, and D. E. Veron. 2009*a*. Winter Northern Hemisphere weather patterns remember summer Arctic sea-ice extent. Geophysical Research Letters 36:L07503.

Francis, J. A., and S. J. Vavrus. 2012. Evidence linking Arctic amplification to extreme weather in mid-latitudes. Geophysical Research Letters 39:L06801.

Francis, J. A., D. M. White, J. J. Cassano, W. J. Gutowski, Jr., L. D. Hinzman, M. M. Holland, M. A. Steele, and C. J. Vörösmarty. 2009*b*. An Arctic hydrologic system in transition: feedbacks and impacts on terrestrial, marine, and human life. Journal of Geophysical Research 114:G04019.

Frauenfeld, O. W., T. Zhang, R. G. Barry, and D. Gilichinsky. 2004. Interdecadal changes in seasonal freeze and thaw depths in Russia. Journal of Geophysical Research 109:D05101.

Gabric, A. J., B. Qu, P. Matrai, and A. C. Hirst. 2005. The simulated response of dimethylsulfide production in the Arctic Ocean to global warming. Tellus A and B 57B:391–403.

Gardner, A. S., G. Moholdt, B. Wouters, G. J. Wolken, D. O. Burgess, M. J. Sharp, J. G. Cogley, C. Braun, and C. Labine. 2011. Sharply increased mass loss from glaciers and ice caps in the Canadian Arctic Archipelago. Nature 473:357–360.

Ghatak, D., A. Frei, G. Gong, J. Stroeve, and D. Robinson. 2010. On the emergence of an Arctic amplification signal in terrestrial Arctic snow extent. Journal of Geophysical Research 115:D24105.

Giles, K. A., S. W. Laxon, A. L. Ridout, D. J. Wingham, and S. Bacon. 2012. Western Arctic Ocean freshwater storage increased by wind-driven spin-up of the Beaufort Gyre. Nature Geoscience 5:194–197.

Graversen, R. G. 2006. Do changes in the midlatitude circulation have any impact on the Arctic surface air temperature trend? Journal of Climate 19:5422–5438.

Graversen, R. G., T. Mauritsen, M. Tjernstrom, E. Kallen, and G. Svensson. 2008. Vertical structure of recent Arctic warming. Nature 451:53–56.

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

Grebmeier, J. M., S. E. Moore, J. E. Overland, K. E. Frey, and R. Gradinger. 2010. Biological response to recent Pacific Arctic sea ice retreats. EOS 91:161–168.

Grebmeier, J. M., J. E. Overland, S. E. Moore, E. V. Farley, E. Carmack, L. W. Cooper, K. E. Frey, J. H. Helle, F. McLaughlin, and S. L. McNutt. 2006. A major ecosystem shift in the Northern Bering Sea. Science 311:1461–1464.

Grosse, G., L. Schirrmeister, C. Siegert, V. Kunitsky, E. A. Slagoda, A. A. Andreev, and A. Y. Dereviagyn. 2007. Geological and geomorphological evolution of a sedimentary periglacial landscape in northeast Siberia during the Late Quaternary. Geomorphology 86:25–51.

Haeberli, W., I. Gärtner-Roer, M. Hoelzle, F. Paul, and M. Zemp, editors. 2009. Glacier Mass Balance Bulletin No. 10 (2006–2007). World Glacier Monitoring Service, Zurich, Switzerland.

Hanna, E., J. Cappelen, R. Allan, T. Jónsson, F. le Blancq, T. Lillington, and K. Hickey. 2008*b*. New insights into North European and North Atlantic surface pressure variability, storminess and related climatic change since 1830. Journal of Climate 21:6739–6766.

Hanna, E., P. Huybrechts, K. Steffen, J. Cappelen, R. Huff, C. Shuman, T. Irvine-Fynn, S. Wise, and M. Griffiths. 2008*a*. Increased runoff from melt from the Greenland Ice Sheet: a response to global warming. Journal of Climate 21(2):331–341.

Hanna, E., J. M. Jones, J. Cappelen, S. H. Mernild, L. Wood, K. Steffen, and P. Huybrechts. 2012. The influence of North Atlantic atmospheric and oceanic forcing effects on 1900–2010 Greenland summer climate and ice melt/runoff. International Journal of Climatology 33:862–880.

Hayes, D. J., A. D. McGuire, D. W. Kicklighter, K. R. Gurney, T. J. Burnside, and J. M. Melillo. 2011. Is the northern high latitude land-based CO_2 sink weakening? Global Biogeochemical Cycles 25:GB3018.

Hegseth, E. N., and A. Sundfjord. 2008. Intrusion and blooming of Atlantic phytoplankton species in the high Arctic. Journal of Marine Systems 74:108–119.

Hinzman, L. D., et al. 2005. Evidence and implications of recent climate change in northern Alaska and other Arctic regions. Climatic Change 72(3):251–298.

Hinzman, L. D., W. R. Bolton, K. C. Petrone, J. B. Jones, Jr., and P. C. Adams. 2006*a*. Watershed hydrology and chemistry in the Alaskan boreal forest: the central role of permafrost. Pages 269–284 *in* F. S. Chapin, III, M. W. Oswood, K. Van Cleve, L. A. Viereck, and D. L. Verbyla, editors. Alaska's changing boreal forest. Oxford University Press, New York, New York, USA.

Hinzman, L., M. Fukuda, D. V. Sandberg, F. S. Chapin, III, and D. Dash. 2003*a*. FROSTFIRE: an experimental approach to predicting the climate feedbacks from the changing boreal fire regime. Journal of Geophysical Research 108(D1):8153.

Hinzman, L. D., R. E. Gieck, and D. L. Kane. 2008. Spatial and temporal variation of soil temperatures and Arctic hydrology in the Kuparuk River Basin, Alaska. Pages 711–716 *in* D. L. Kane and K. M. Hinkel, editors. University of Alaska, Institute of Northern Engineering, Fairbanks, Alaska, USA.

Hinzman, L. D., D. J. Goering, and D. L. Kane. 1998. A distributed thermal model for calculating soil temperature profiles and depth of thaw in permafrost regions. Journal of Geophysical Research 103(D22):28, 975–992.

Hinzman, L. D., D. L. Kane, and K. Yoshikawa. 2003*b*. Soil moisture response to a changing climate in Arctic regions. Tôhoku Geophysical Journal 36(4):369–373.

Hinzman, L. D., D. L. Kane, K. Yoshikawa, A. Carr, W. R. Bolton, and M. Fraver. 2003*c*. Hydrological variations among watersheds with varying degrees of permafrost. Pages 407–411 *in* Proceedings of 8th International Conference on Permafrost, Zurich, Switzerland, 21–25 July 2003. Aa Balkema, Leiden, The Netherlands.

Hinzman, L. D., L. A. Viereck, P. C. Adams, V. E. Romanovsky, and K. Yoshikawa. 2006*b*. Climate and permafrost dynamics of the Alaskan boreal forest. Pages 39–61 *in* F. S. Chapin, III, M. W. Oswood, K. Van Cleve, L. A. Viereck, and D. L. Verbyla, editors. Alaska's changing boreal forest. Oxford University Press, New York, New York, USA.

Houghton, J. T., Y. Ding, D. J. Griggs, M. Noguer, P. J. v. d. Linden, X. Dai, K. Maskell, and C. A. Johnson, editors. 2001. Climate change 2001: the scientific basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, UK.

Howat, I. M., and A. Eddy. 2011. Multi-decadal retreat of Greenland's marine-terminating glaciers. Journal of Glaciology 57(203):389–396.

Hunt, G. L., P. Stabeno, G. Walters, E. Sinclair, R. D. Brodeur, J. M. Napp, and N. A. Bond. 2002. Climate change and control of the southeastern Bering Sea pelagic ecosystem. Deep-Sea Research Part II: Topical Studies in Oceanography 49(26):5821–5853.

Ivanov, V., and E. Watanabe. 2013. Does Arctic sea ice reduction foster shelf–basin exchange? Ecological Applications 23:1765–1777.

Jin, M., C. J. Deal, S. Lee, S. Elliott, E. Hunke, M. Maltrud, and N. Jeffery. 2012. Investigation of Arctic sea ice and ocean primary production for the period 1992 to 2007 using a 3-D global ice–ocean ecosystem model. Deep-Sea Research II 81–84(2012):28–35.

Jin, M., C. J. Deal, J. Wang, and C. P. McRoy. 2009. Response of lower trophic level production to long-term climate change in the southeastern Bering Sea. Journal of Geophysical Research 114:C04010.

Jones, B. M., C. D. Arp, M. T. Jorgenson, K. M. Hinkel, J. A. Schmutz, and P. L. Flint. 2009. Increase in the rate and uniformity of coastline erosion in Arctic Alaska. Geophysical Research Letters 36:L03503.

Jorgenson, M. T., C. H. Racine, J. C. Walters, and T. E. Osterkamp. 2001. Permafrost degradation and ecological changes associated with a warming climate in central Alaska. Climatic Change 48(4):551–571.

Jorgenson, M. T., Y. L. Shur, and E. R. Pullman. 2006. Abrupt increase in permafrost degradation in Arctic Alaska. Geophysical Research Letters 33:L02503.

Joughin, I., S. B. Das, M. A. King, B. E. Smith, I. M. Howat, and T. Moon. 2008. Seasonal speedup along the western flank of the Greenland Ice Sheet. Science 320:781–783.

Juntunen, V., S. Neuvonen, Y. Norokorpi, and T. Tasanen. 2002. Potential for timberline advance in northern Finland, as revealed by monitoring during 1983–1999. Arctic 55:348–361.

Kane, D. L., K. M. Hinkel, D. J. Goering, L. D. Hinzman, and S. I. Outcalt. 2001. Non-conductive heat transfer associated with frozen soils. Global and Planetary Change 29:275–292.

Kane, D. L., L. D. Hinzman, J. P. McNamara, Z. Zhang, and C. S. Benson. 2000. An overview of a nested watershed study in Arctic Alaska. Nordic Hydrology 4/5:245–266.

Kargel, J. S., et al. 2011. Brief communication: Greenland's shrinking ice cover: "fast times" but not that fast. Cryosphere Discussion 5:3207–3219.

Kasischke, E. S., and M. R. Turetsky. 2006. Recent changes in the fire regime across the North American boreal region. Geophysical Research Letters 33:L09703.

Kasischke, E. S., et al. 2010. Alaska's changing fire regime—implications for the vulnerability of its boreal forests. Canadian Journal of Forest Research 40:1313–1324.

Kattsov, V. M., V. E. Ryabinin, J. E. Overland, M. C. Serreze, M. Visbeck, J. E. Walsh, and X. Zhang. 2010. Arctic sea ice change: a grand challenge of climate science. Journal of Glaciology 56(200):1115–1121.

Kay, J. E., M. M. Holland, and A. Jahn. 2011. Inter-annual to multi-decadal Arctic sea ice extent trends in a warming world. Geophysical Research Letters 38:L15708.

Kharuk, V. I., M. L. Dvinskaya, and K. J. Ranson. 2005*a*. The spatiotemporal pattern of fires in northern taiga larch forests of central Siberia. Russian Journal of Ecology 36:302–311.

Kharuk, V. I., M. L. Dvinskaya, K. J. Ranson, and S. T. Im. 2005*b*. Expansion of evergreen conifers to the larch-dominated zone and climatic trends. Russian Journal of Ecology 36:186–193.

Kicklighter, D. W., D. J. Hayes, J. W. McClelland, B. J. Peterson, A. D. McGuire, and J. M. Melillo. 2013. Insights and issues with simulating terrestrial DOC loading of Arctic river networks. Ecological Applications 23:1817–1836.

Koven, C. D., B. Ringeval, P. Friedlingstein, P. Ciais, P. Cadule, D. Khvorostyanov, G. Krinner, and C. Tarnocai. 2011. Permafrost carbon-climate feedbacks accelerate global warming. Proceedings of the National Academy of Sciences USA 108:14769–14774.

Kwok, R. 2009. Outflow of Arctic sea ice into the Greenland and Barents seas: 1979–2007. Journal of Climate 22:2438–2457.

Kwok, R., G. F. Cunningham, M. Wensnahan, I. Rigor, H. J. Zwally, and D. Yi. 2009. Thinning and volume loss of the Arctic Ocean sea ice cover: 2003–2008. Journal of Geophysical Research 114:C07005.

Kwok, R., and D. A. Rothrock. 2009. Decline in Arctic sea ice thickness from submarine and ICESat records: 1958–2008. Geophysical Research Letters 36:L15501.

Lachenbruch, A. H., and B. V. Marshall. 1986. Changing climate: geothermal evidence from permafrost in the Alaskan Arctic. Science 234:689–696.

Lammers, R. B., A. I. Shiklomanov, C. J. Vörösmarty B. M. Fekete, and B. J. Peterson. 2001. Assessment of contemporary Arctic river runoff based on observational discharge records. Journal of Geophysical Research 10:63321–3334.

Lavoie, C., and S. Payette. 1994. Recent fluctuations of the lichen–spruce forest limit in subArctic Quebec. Journal of Ecology 82:725–734.

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.

Leck, C., and E. Bigg. 2005. Source and evolution of the marine aerosol—a new perspective. Geophysical Research Letters 32:L19803.

Leclercq, P. W., J. Oerlemans, and J. G. Cogley. 2011. Estimating the glacier contribution to sea-level rise for the period 1800–2005. Surveys in Geophysics 32:519–535.

Lee, S. H., H. M. Joo, Z. Liu, J. Chen, and J. He. 2012. Phytoplankton productivity in newly opened waters of the Western Arctic Ocean. Deep-Sea Research Part II: Topical Studies in Oceanography Part II 81:18–27.

Lengaigne, M., G. Madec, L. Bopp, C. Menkes, O. Aumont, and P. Cadule. 2009. Bio-physical feedbacks in the Arctic Ocean using an Earth system model. Geophysical Research Letters 36:L21602.

Li, W. K. W., F. A. McLaughlin, C. Lovejoy, and E. C. Carmack. 2009. Smallest algae thrive as the Arctic Ocean freshens. Science 326:539.

Lindsay, R. W., J. Zhang, A. Schweiger, M. A. Steele, and H. Stern. 2009. Arctic sea ice retreat in 2007 follows thinning trend. Journal of Climate 22:165–176.

Liston, G. E., and C. A. Hiemstra. 2011. The changing cryosphere: pan-Arctic snow trends (1979–2009). Journal of Climate 24:5691–5712.

Liu, J., J. A. Curry, H. Wang, M. Song, and R. M. Horton. 2012. Impact of declining Arctic sea ice on winter snowfall. Proceedings of the National Academy of Sciences USA 109:4074–4079.

Lloyd, A. H., T. S. Rupp, C. L. Fastie, and A. M. Starfield. 2003. Patterns and dynamics of treeline advance in the Seward Peninsula, Alaska. Journal of Geophysical Research 108(D2):8161.

Mantua, N. J., and S. R. Hare. 2002. The Pacific Decadal Oscillation. Journal of Oceanography 58:35–44.

Mathis, J. T., J. N. Cross, and N. R. Bates. 2011. Coupling primary production and terrestrial runoff to ocean acidification and carbonate mineral suppression in the eastern Bering Sea. Journal of Geophysical Research 116:C02030.

Mazhitova, G. G., G. Malkova, O. Chestnykh, and D. Zamolodchikov. 2008. Recent decade thaw depth dynamics in the European Russian Arctic based on the Circumpolar Active Layer Monitoring (CALM) data. Pages 1155–1160 *in* D. L. Kane and K. M. Hinkel, editors. Ninth International Conference on Permafrost Proceedings. Institute of Northern Engineering, University of Alaska, Fairbanks, USA.

McClelland, J. W., R. M. Holmes, B. J. Peterson, and M. Stieglitz. 2004. Increasing river discharge in the Eurasian Arctic: consideration of dams, permafrost thaw, and fires as potential agents of change. Journal of Geophysical Research 109:D18102.

McDonald, K., J. Kimball, E. Njoku, R. Zimmermann, and M. Zhao. 2004. Variability in springtime thaw in the terrestrial high latitudes: Monitoring a major control on the biospheric assimilation of atmospheric CO_2 with spaceborne microwave remote sensing. Earth Interactions 8(20):1–23.

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

McGuire, A. D., et al. 2004. Land cover disturbances and feedbacks to the climate system in Canada and Alaska. Pages 139–161 *in* G. Gutman, A. C. Janetos, C. O. Justice, E. F. Moran, J. F. Mustard, R. R. Rindfuss, D. Skole, B. L. Turner II, and M. A. Cochrane, editors. Land change science: observing, monitoring, and understanding trajectories of change on the Earth's surface. Kluwer Academic Publishers, Dordrecht, The Netherlands.

McGuire, A. D., et al. 2010. An analysis of the carbon balance of the Arctic Basin from 1997 to 2006. Tellus 62B:455–474.

McGuire, A. D., et al. 2012. An assessment of the carbon balance of Arctic tundra: comparisons among observations, process models, and atmospheric inversions. Biogeosciences Discussions 9:4543–5494.

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

Meier, M. F., M. B. Dyurgerov, U. K. Rick, S. O'Neel, W. T. Pfeffer, R. S. Anderson, S. P. Anderseon, and A. F. Glazovsky. 2007. Glaciers dominate eustatic sea-level rise in the 21st century. Science 317:1064–1067.

Mernild, S. H., N. T. Knudsen, W. H. Lipscomb, J. C. Yde, J. K. Malmros, B. H. Jakobsen, and B. Hasholt. 2011*a*. Increasing mass loss from Greenland's Mittivakkat Gletscher. Cryosphere 5:341–348.

Mernild, S. H., N. T. Knudsen, J. C. Yde, M. J. Hoffman, W. L. Lipscomb, R. S. Fausto, E. Hanna, and J. K. Malmros. 2013*a*. Volume and velocity changes at Mittivakkat Gletscher, Southeast Greenland, 1994–2012. Journal of Glaciology 59:660–670.

Mernild, S. H., W. H. Lipscomb, D. B. Bahr, V. Radić, and M. Zemp. 2013*b*. Global glacier retreat: a revised assessment of committed mass losses and sampling uncertainties. Cryosphere Discussion 7:1–19.

Mernild, S. H., and G. E. Liston. 2012. Greenland freshwater runoff. Part II: distribution and trends, 1960–2010. Spatial runoff distribution and trends from Greenland to adjacent seas, 1960–2010. Journal of Climate 25(17):6015–6035.

Mernild, S. H., G. E. Liston, C. A. Hiemstra, and J. H. Christensen. 2010. Greenland Ice Sheet surface mass-balance modeling in a 131-year perspective 1950–2080. Journal of Hydrometeorology 11(1):3–25.

Mernild, S. H., J. K. Malmros, J. C. Yde, and N. T. Knudsen. 2012. Multi-decadal marine and land-terminating glacier retreat in Ammassalik region, Southeast Greenland. Cryosphere 6:625–639.

Mernild, S. H., T. Mote, and G. E. Liston. 2011*b*. Greenland Ice Sheet surface melt extent and trends, 1960–2010. Journal of Glaciology 57(204):621–628.

Morison, J., R. Kwok, C. Peralta-Ferriz, M. Alkire, I. Rigor, R. Andersen, and M. Steele. 2012. Changing Arctic Ocean freshwater pathways. Nature 481:66–70.

Morison, J., J. Wahr, R. Kwok, and C. Peralta-Ferriz. 2007. Recent trends in Arctic Ocean mass distribution revealed by GRACE. Geophysical Research Letters 34:L07602.

Mote, T. L. 2007. Greenland surface melt trends 1973–2007: evidence of a large increase in 2007. Geophysical Research Letters 34(22):L22507.

Mueter, F. J., and M. A. Litzow. 2008. Sea ice retreat alters the biogeography of the Bering Sea continental shelf. Ecological Applications 18:309–320.

Myers-Smith, I. H., et al. 2011. Shrub expansion in tundra ecosystems: dynamics, impacts, and research priorities. Environmental Research Letters 6:045509.

Nerem, R. S., D. P. Chambers, C. Choe, and G. T. Mitchum. 2010. Estimating mean sea level change from the TOPEX and Jason altimeter mission. Marine Geodesy 33:435–466.

Nghiem, S. V., I. G. Rigor, D. K. Perovich, P. Clemente-Colon, J. W. Weatherly, and G. Neumann. 2007. Rapid reduction of Arctic perennial sea ice. Geophysical Research Letters 34:L19504.

Nick, F. M., A. Vieli, I. M. Howat, and I. Joughin. 2009. Large-scale changes in Greenland outlet glacier dynamics triggered at the terminus. Nature Geoscience 2(2):110–114.

NSIDC. 2010. Weather and feedbacks lead to third-lowest extent. National Snow and Ice Data Center, University of Colorado, Boulder, Colorado, USA. http://nsidc.org/arcticseaicenews/2010/10/weather-and-feedbacks-lead-to-third-lowest-extent/

NSIDC. 2011. Summer 2011: Arctic sea ice near record lows. National Snow and Ice Data Center, University of Colorado, Boulder, Colorado, USA. http://nsidc.org/arcticseaicenews/2011/10/summer-2011-arctic-sea-ice-near-record-lows/

Oechel, W. C., G. L. Vourlitis, S. J. Hastings, R. C. Zulueta, L. D. Hinzman, and D. L. Kane. 2000. Acclimation of ecosystem CO_2 exchange in the Alaskan Arctic in response to decadal climate warming. Nature 406:978–981.

Ogi, M., I. G. Rigor, M. G. McPhee, and J. M. Wallace. 2008. Summer retreat of Arctic sea ice: role of summer winds. Geophysical Research Letters 35:L23701.

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.

Osterkamp, T. E. 2008. Thermal state of permafrost in Alaska during the fourth quarter of the twentieth century (plenary paper). Pages 1333–1338 *in* D. L. Kane and K. M. Hinkel, editors. Proceedings of the Ninth International Conference on Permafrost. Institute of Northern Engineering, University of Alaska, Fairbanks, Alaska, USA.

Overduin, P. P., H.-W. Hubberten, V. Rachold, N. Romanovskii, M. Grigoriev, and M. Kasymskaya. 2007. The evolution and degradation of coastal and offshore permafrost in the Laptev and East Siberian seas during the last climatic cycle. Geological Society of America, Special Papers 426:97–111.

Overduin, P. P., and D. L. Kane. 2006. Frost boils and soil ice content: field observations. Permafrost and Periglacial Processes 17:291–307.

Overland, J. E., and M. Wang. 2010. Large-scale atmospheric circulation changes associated with the recent loss of Arctic sea ice. Tellus 62A:1–9.

Overland, J. E., M. Wang, and S. Salo. 2008. The recent Arctic warm period. Tellus 60A:589–597.

Pavlov, A. V., and N. G. Moskalenko. 2002. The thermal regime of soils in the north of western Siberia. Permafrost and Periglacial Processes 13:43–51.

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

Peterson, B. J., R. M. Holmes, J. W. McClelland, C. J. Vorosmarty, I. A. Shiklomanov, A. I. Shiklomanov, R. B. Lammers, and S. Rahmstorf. 2002. Increasing river discharge to the Arctic Ocean. Science 298:2171–2173.

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

Petrone, K. C., L. D. Hinzman, and R. D. Boone. 2000. Nitrogen and carbon dynamics of storm runoff in three sub-Arctic streams. Pages 167–172 *in* D. L. Kane, editor. Water resources in extreme environments. American Water Resources Association, Middleburg, Virginia, USA.

Péwé, T. L. 1973. Ice wedge casts and past permafrost distribution in North America. Geoforum 4(3):15–26.

Piao, S., X. Wang, P. Ciais, B. Zhu, T. Wang, and J. Liu. 2011. Changes in satellite-derived vegetation growth trend in temperate and boreal Eurasia from 1982–2006. Global Change Biology 17:3228–3239.

Pnyushkov, A., and I. V. Polyakov. 2012. Observations of tidally-induced currents over the continental slope of the Laptev Sea, Arctic Ocean. Journal of Physical Oceanography 42:78–94.

Polyakov, I. V., G. V. Alekseev, L. A. Timokhov, U. Bhatt, R. L. Colony, H. L. Simmons, D. Walsh, J. E. Walsh, and V. F. Zakharov. 2004. Variability of the intermediate Atlantic Water of the Arctic Ocean over the last 100 years. Journal of Climate 17(23):4485–4497.

Polyakov, I. V., V. Alexeev, G. I. Belchansky, I. A. Dmitrenko, V. Ivanov, S. Kirillov, A. Korablev, M. Steele, L. A. Timokhov, and I. Yashayaev. 2008. Arctic Ocean freshwater changes over the past 100 years and their causes. Journal of Climate 21(2):364–384.

Polyakov, I. V., et al. 2005. One more step toward a warmer Arctic. Geophysical Research Letters 32:L17605.

Polyakov, I. V., et al. 2011. Fate of early-2000's Arctic warm water pulse. Bulletin of the American Meteorological Society 92:561–566.

Polyakov, I. V., R. Kwok, and J. E. Walsh. 2012. Recent changes of Arctic multiyear sea-ice coverage and their likely causes. Bulletin of American Meteorological Society 93(2):145–151.

Polyakov, I. V., et al. 2010. Arctic Ocean warming reduces polar ice cap. Journal of Physical Oceanography 40:2743–2756.

Popova, E., A. Yool, A. Coward, F. Dupont, C. Deal, S. Elliott, E. Hunke, M. Jin, M. Steele, and J. Zhang. 2012. What controls primary production in the Arctic Ocean? Results from an intercomparison of five general circulation models with biogeochemistry. Journal of Geophysical Research: Oceans 117:C00D12.

Porter, D. F., J. J. Cassano, and M. C. Serreze. 2012. Local and large-scale atmospheric responses to reduced Arctic sea ice and ocean warming in the WRF model, Journal of Geophysical Research 117:D11115.

Proshutinsky, A., R. Krishfield, M.-L. Timmermans, J. Toole, E. Carmack, F. McLaughlin, W. J. Williams, S. Zimmermann, M. Itoh, and K. Shimada. 2009. Beaufort Gyre freshwater reservoir: State and variability from observations. Journal of Geophysical Research 114:C00A10.

Rachold, V., M. N. Grigoriev, F. E. Are, S. M. Solomon, E. Reimnitz, H. Kassens, and M. Antonov. 2000. Coastal erosion vs. riverine sediment discharge in the Arctic shelf seas. International Journal of Earth Sciences 89:450–460.

Radić, V., and R. Hock. 2010. Regional and global volumes of glaciers derived from statistical upscaling of glacier inventory data. Journal of Geophysical Research 115:F01010.

Radić, V., and R. Hock. 2011. Regionally differentiated contribution of mountain glaciers and ice caps to future sea-level rise. Nature Geoscience 4:91–94.

Rahmstorf, S. 1995. Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. Nature 378:145–149.

Rahmstorf, S., et al. 2005. Thermohaline circulation hysteresis: a model intercomparison. Geophysical Research Letters 32:L23605.

Randerson, J. T., et al. 2006. The impact of boreal forest fire on climate warming. Science 314(5802):1130–1132.

Rawlins, M. A., et al. 2010. Analysis of the Arctic system for freshwater cycle intensification: observations and expectations. Journal of Climate 23:5715–5737.

Retamal, L., S. Bonilla, and W. F. Vincent. 2008. Optical gradients and phytoplankton production in the Mackenzie River and the coastal Beaufort Sea. Polar Biology 31:363–379.

Rignot, E., J. E. Box, E. Burgess, and E. Hanna. 2008. Mass balance of the Greenland ice sheet from 1958–2007. Geophysical Research Letters 35:L20502.

Rignot, E., and P. Kanagaratnam. 2006. Changes in the velocity structure of the Greenland Ice Sheet. Science 311(5673):986–990.

Roach, J., B. Griffith, D. Verbyla, and J. Jones. 2011. Mechanisms influencing changes in lake area in Alaskan boreal forest. Global Change Biology 17:2567–2583.

Roberts, A., L. D. Hinzman, J. E. Walsh, M. Holland, J. Cassano, R. Döscher, H. Mitsudera, and A. Sumi. 2010. A science plan for regional Arctic system modeling, a report to the National Science Foundation from the International Arctic Science Community. International Arctic Research Center Technical Papers 10-0001. International Arctic Research Center, University of Alaska, Fairbanks, Alaska, USA.

Romanovsky, V. E., et al. 2010*a*: Thermal state of permafrost in Russia. Permafrost and Periglacial Processes 21:136–155.

Romanovsky, V. E., S. L. Smith, and H. H. Christiansen. 2010*b*. Permafrost thermal state in the polar Northern Hemisphere during the International Polar Year 2007–2009: a synthesis. Permafrost and Periglacial Processes 21:106–116.

Rothrock, D. A., Y. Yu, and G. A. Maykut. 1999. Thinning of the Arctic sea ice cover. Geophysical Research Letters 26:3469–3472.

Saito, K., T. Zhang, D. Yang, S. Marchenko, R. G. Barry, V. Romanovsky, and L. D. Hinzman. 2013. Influence of the physical terrestrial Arctic in the eco-climate system. Ecological Applications 23:1778–1797.

Sarmiento, J. L., and N. Gruber. 2006. Ocean biogeochemical dynamics. Princeton University Press, Princeton, New Jersey, USA.

Schaefer, K., H. Lantuit, V. E. Romanovsky, and E. A. G. Schuur 2012. Policy implications of warming permafrost. United Nations Environment Programme, Washington, D.C., USA.

Schaefer, K., T. Zhang, L. Bruhwiler, and A. P. Barrett. 2011. Amount and timing of permafrost carbon release in response to climate warming. Tellus 63B:165–180.

Schauer, U., E. Fahrbach, S. Osterhus, and G. Rohardt. 2004. Arctic warming through the Fram Strait: oceanic heat transport from 3 years of measurements. Journal of Geophysical Research 109:C06026.

Schneider von Deimling, T., M. Meinshausen, A. Levermann, V. Huber, K. Frieler, D. M. Lawrence, and V. Brovkin. 2012. Estimating the permafrost carbon feedback on global warming. Biogeosciences 9:649–665.

Schuur, E. A. G., et al. 2008. Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. BioScience 58:701–714.

Schuur, E. A. G., et al. 2011. High risk of permafrost thaw. Nature 480:32–33.

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

Schweiger, A. J., J. Zhang, R. W. Lindsay, and M. Steele. 2008. Did unusually sunny skies help drive the sea ice anomaly of 2007. Geophysical Research Letters 35:L10503.

Semiletov, I. P., I. I. Pipko, N. E. Shakhova, O. V. Dudarev, S. P. Pugach, A. N. Charkin, C. P. McRoy, D. Kosmach, and O. Gustafsson. 2011. On the biogeochemical signature of the Lena River from its headwaters to the Arctic Ocean. Biogeosciences Discuss 8:2093–2143.

Serreze, M. C., A. P. Barrett, and J. C. Stroeve. 2012. Recent changes in tropospheric water vapor over the Arctic as assessed from radiosondes and atmospheric reanalyses. Journal of Geophysical Research 117:D10104.

Serreze, M. C., A. P. Barrett, J. C. Stroeve, D. N. Kindig, and M. M. Holland. 2009. The emergence of surface-based Arctic amplification. Cryosphere 3:11–19.

Serreze, M. C., and A. J. Etringer. 2003. Precipitation characteristics of the Eurasian Arctic drainage system. International Journal of Climatology 23:1267–1291.

Serreze, M. C., and J. A. Francis. 2006. The Arctic amplification debate. Climate Change 76:241–264.

Serreze, M. C., J. E. Walsh, F. S. Chapin, III, T. Osterkamp, M. Dyurgerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry. 2000. Observational evidence of recent change in the northern high latitude environment. Climate Change 46:159–207.

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

Shimada, K., T. Kamoshida, M. Itoh, S. Nishino, E. Carmack, F. McLaughlin, S. Zimmermann, and A. Proshutinsky. 2006. Pacific Ocean inflow: influence on catastrophic reduction of sea ice cover in the Arctic Ocean. Geophysical Research Letters 33:L08605.

Skific, N., J. A. Francis, and J. J. Cassano. 2009. Attribution of projected changes in atmospheric moisture transport in the Arctic: a self-organizing map perspective. Journal of Climate 22(15):4135–4153.

Smedsrud, L. H., A. Sorteberg, and K. Kloster. 2008. Recent and future changes of the Arctic sea-ice cover, Geophysical Research Letters 35:L20503.

Smith, L. C., Y. Sheng, and G. M. MacDonald. 2007. A first pan-Arctic assessment of the influence of glaciation, permafrost, topography and peatlands on northern lake distribution. Permafrost and Periglacial Processes 18:201–208.

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

Smith, S. L., M. M. Burgess, D. Riseborough, and F. M. Nixon. 2005*b*. Recent trends from Canadian permafrost thermal monitoring network sites. Permafrost and Periglacial Processes 16:19–30.

Smith, S. L., V. E. Romanovsky, A. G. Lewkowicz, C. R. Burn, M. Allard, G. D. Clow, K. Yoshikawa, and J. Throop. 2010. Thermal state of permafrost in North America—a contribution to the International Polar Year. Permafrost and Periglacial Processes 21:117–135.

Smol, J. P., and M. S. V. Douglas. 2007. Crossing the final ecological threshold in high Arctic ponds. Proceedings of the National Academy of Sciences USA 104:12395–12397.

Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, M. Tignor, and H. L. Miller, editors. 2007. Summary for policymakers *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. Cambridge University Press, Cambridge, UK.

Steffen, K. 1995. Surface energy exchange at the equilibrium line on the Greenland ice sheet during onset of melt. Annals of Glaciology 21:13–18.

Steffen, K., P. U. Clark, J. G. Cogley, D. Holland, S. Marshall, E. Rignot, and R. Thomas. 2008. Rapid changes in glaciers and ice sheets and their impacts on sea level. Pages 29–66 *in* Abrupt climate change. US Geological Survey, Reston, Virginia, USA.

Steinacher, M., F. Joos, T. L. Frolicher, G.-K. Platter, and S. C. Doney. 2009. Imminent ocean acidification of the Arctic projected with the NCAR global coupled carbon-cycle climate model. Biogeosciences 6:515–533.

Steinacher, M., et al. 2010. Projected 21st century decrease in marine productivity: a multi-model analysis. Biogeosciences 7:979–1005.

Stieglitz, M., S. J. Déry, V. E. Romanovsky, and T. E. Osterkamp. 2003. The role of snow cover in the warming of Arctic permafrost. Geophysical Research Letters 30(13):1721.

Stroeve, J. C., J. Maslanik, M. C. Serreze, I. Rigor, W. Meier, and C. Fowler. 2011. Sea ice response to an extreme negative phase of the Arctic Oscillation during winter 2009/2010. Geophysical Research Letters 38:L02502.

Stroeve, J. C., M. C. Serreze, J. E. Kay, M. M. Holland, W. N. Meier, and A. P. Barrett. 2012. The Arctic's rapidly shrinking sea ice cover: a research synthesis. Climate Change 110:1005–1027.

Sturm, M., J. P. McFadden, G. E. Liston, F. S. Chapin, C. H. Racine, and J. Holmgren. 2001. Shrub-snow interactions in Arctic tundra: a hypothesis with climatic implications. Journal of Climate 14:336–344.

Sturm, M., D. K. Perovich, and J. Holmgren. 2002. Thermal conductivity and heat transfer through the snow on the ice of the Beaufort Sea. Journal of Geophysical Research 107(C10):8043.

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.

Tedesco, M. 2007. A new record in 2007 for melting in Greenland. EOS 88(39):383.

Tedesco, M., X. Fettweis, M. R. van den Broeke, R. S. W. van de Wal, C. J. P. P. Smeets, W. J. van de Berg, M. C. Serreze, and J. E. Box. 2011. The role of albedo and accumulation in the 2010 melting record in Greenland. Environmental Research Letters 6:014005.

Thompson, D. W. J., and J. M. Wallace. 2000. Annular modes in the extratropical circulation. Part I: month-to-month variability. Journal of Climate 13:1000–1016.

Timmermans, M.-L., A. Proshutinsky, R. A. Krishfield, D. K. Perovich, J. A. Richter-Menge, T. P. Stanton, and J. M. Toole. 2011. Surface freshening in the Arctic Ocean's Eurasian Basin: an apparent consequence of recent change in the wind-driven circulation. Journal of Geophysical Research 116:C00D03.

Toole, J. M., M.-L. Timmermans, D. K. Perovich, R. A. Krishfield, A. Proshutinsky, and J. A. Richter-Menge. 2010. Influences of the ocean surface mixed layer and thermohaline stratification on Arctic sea ice in the central Canada Basin. Journal of Geophysical Research 115:C10018.

Tremblay, J.-E., and J. Gagnon. 2009. The effect of irradiance and nutrient supply on the productivity of Arctic waters: a perspective on climate change. Pages 73–94 *in* J. C. J. Nihpul and A. G. Kostianoy, editors. Influence of climate change on the changing Arctic and sub-Arctic conditions. Springer, New York, New York, USA.

Trochim, E. D., A. Prakash, and D. L. Kane. 2008. Delineating water tracks in headwater drainage basins of the Alaskan Arctic. Fall Meeting 2008, abstract #C31E-0558. American Geophysical Union, Washington, D.C., USA.

Troy, T. J., J. Sheffield, and E. F. Wood. 2012. The role of winter precipitation and temperature on northern Eurasian streamflow trends. Journal of Geophysical Research 117:D05131.

Ueyama, M., H. Iwata, Y. Harazono, E. S. Euskirchen, W. C. Oechel, and D. Zona. 2013. Growing season and spatial variations of carbon fluxes of Arctic and boreal ecosystems in Alaska (USA). Ecological Applications. 23:1798–1816.

U.S. Arctic Research Commission. 2010. Report on goals and objectives for Arctic research 2009–2010. U.S. Arctic Research Program Plan. U.S. Arctic Research Commission, Arlington, Virginia, USA.

van den Broeke, M. R., J. L. Bamber, J. Ettema, E. Rignot, E. Schrama, W. J. van de Berg, E. van Meijgaard, I. Velicogna, and B. Wouters. 2009. Partitioning recent Greenland mass loss. Science 326:984–986.

van der Wal, R. S. W., W. Boot, M. R. van den Broeke, C. J. P. P. Smeets, C. H. Reijmer, J. J. A. Donker, and J. Oerlemans. 2008. Large and rapid melt-induced velocity changes in the ablation zone of the Greenland Ice Sheet. Science 321:111–113.

van Huissteden, J., C. Berrittella, F. J. W. Parmentier, Y. Mi, T. C. Maximov, and A. J. Dolman. 2011. Methane emissions from permafrost thaw lakes limited by lake drainage. Nature Climate Change 1:119–123.

Vlassova, T. K. 2002. Human impacts on the tundra-taiga zone dynamics: the case of the Russian lesotundra. Pages 30–36 *in* T. V. Callaghan, editor. Dynamics of the tundra–taiga interface. Ambio Special Report 12.

Wahr, J. 2011. Greenland mass changes from GRACE. 2011. Pages 132–133 *in* J. E. Box et al., editors. The Arctic report card, Greenland. NOAA Report 2011. NOAA, Washington, D.C., USA.

Walsh, J. E., J. E. Overland, P. Groisman, and B. Rudolf. 2012. Arctic climate: recent variations. Chapter 2 *in* Snow, water, ice and permafrost in the Arctic (SWIPA): climate change and the cryosphere. Arctic Monitoring and Assessment Programme, Oslo, Norway.

Walter, K. M., S. A. Zimov, J. P. Chanton, D. Verbyla, and F. S. Chapin. 2006. Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming. Nature 443:71–75.

Wang, J., J. Zhang, E. Watanabe, M. Ikeda, K. Mizobata, J. E. Walsh, X. Bai, and B. Wu. 2009. Is the Dipole Anomaly a major driver to record lows in Arctic summer sea ice extent? Geophysical Research Letters 36:L05706.

Wang, X. L., and J. R. Key. 2005. Arctic surface, cloud, and radiative properties based on the AVHRR Polar Pathfinder dataset. Part II. Recent trends. Journal of Climate 18:2575–2593.

Watanabe, E., J. Wang, A. Sumi, and H. Hasumi. 2006. Arctic dipole anomaly and its contribution to sea ice export from the Arctic Ocean in the 20th century. Geophysical Research Letters 33:L23703.

White, D. M., et al. 2007. The Arctic freshwater system: changes and impacts. Journal of Geophysical Research 112:G04S54.

Wirth, C. 2004. Fire regime and tree diversity in boreal and high elevation forests: Implications for biogeochemical cycles. Pages 309–344 *in* M. Scherer-Lorenzen, C. Körner, and E.-D. Schulze, editors. Forest diversity and function.

Ecological studies 176. Springer, New York, New York, USA.

Woo, M. K. 1986. Permafrost hydrology in North America. Atmosphere—Ocean 24(3):201–234.

Woodgate, R. A., T. Weingartner, and R. Lindsay. 2010. The 2007 Bering Strait oceanic heat flux and anomalous Arctic sea-ice retreat. Geophysical Research Letters 37:L01602.

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 sea ice meltwater. Journal of Geophysical Research 114:C00A05.

Yang, D., D. L. Kane, L. Hinzman, X. Zhang, T. Zhang, and H. Ye. 2002. Siberian Lena River hydrologic regime and recent change. Journal of Geophysical Research 107(D23):4694.

Yang, D., D. Kane, Z. Zhang, D. Legates, and B. Goodison. 2005. Bias-corrections of long-term (1973–2004) daily precipitation data over the northern regions. Geophysical Research Letters 32:L19501.

Yi, S., M.-k. Woo, and M. A. Arain. 2007. Impacts of peat and vegetation on permafrost degradation under climate warming. Geophysical Research Letters 34:L16504.

Yoshikawa, K., and L. D. Hinzman. 2003. Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost. Permafrost and Periglacial Processes 14(2):151–160.

Yoshikawa, K., L. D. Hinzman, and P. Gogineni. 2002. Ground temperature and permafrost mapping using an equivalent latitude/elevation model. Journal of Glaciology and Geocryology 24(5):526–531.

Zemp, M., S. U. Nussbaumer, I. Gärtner-Roer, M. Hoelzle, F. Paul, and W. Haeberli, editors. 2011. Glacier mass balance bulletin no. 11 (2008–2009). World Glacier Monitoring Service, Zurich, Switzerland.

Zhang, J., and J. E. Walsh. 2007. Relative impacts of vegetative coverage and leaf area index on climate change in a greener north. Geophysical Research Letters 34:L15703.

Zhang, X., A. Sorteberg, J. Zhang, R. Gerdes, and J. Comiso. 2008. Recent radical shifts in atmospheric circulations and rapid changes in Arctic climate system. Geophysical Research Letters 35:L22701.

Zhuang, Q., J. M. Melillo, M. C. Sarofin, D. W. Kicklighter, A. D. McGuire, B. S. Felzer, A. Sokolov, R. G. Prinn, P. A. Steudler, and S. Hu. 2006. CO_2 and CH_4 exchanges between land ecosystems and the atmosphere in northern high latitudes over the 21st Century. Geophysical Research Letters 33:L17403.

Zimov, S. A., S. P. Davydov, G. M. Zimova, A. I. Davydova, E. A. G. Schuur, K. Dutta, and F. S. Chapin, III. 2006*a*. Permafrost carbon: stock and decomposability of a globally significant carbon pool. Geophysical Research Letters 33:L20502.

Zimov, S. A., E. A. G. Schuur, and F. S. Chapin, III. 2006*b*. Permafrost and the global carbon budget. Science 312:1612–1613.