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The arctic freshwater system: Changes and impacts

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[1] Dramatic changes have been observed in the Arctic over the last century. Many of these involve the storage and cycling of fresh water. On land, precipitation and river discharge, lake abundance and size, glacier area and volume, soil moisture, and a variety of permafrost characteristics have changed. In the ocean, sea ice thickness and areal coverage have decreased and water mass circulation patterns have shifted, changing freshwater pathways and sea ice cover dynamics. Precipitation onto the ocean surface has also changed. Such changes are expected to continue, and perhaps accelerate, in the coming century, enhanced by complex feedbacks between the oceanic, atmospheric, and terrestrial freshwater systems. Change to the arctic freshwater system heralds changes for our global physical and ecological environment as well as human activities in the Arctic. In this paper we review observed changes in the arctic freshwater system over the last century in terrestrial, atmospheric, and oceanic systems.

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1. Introduction

[2] Substantial changes in the Arctic have been observed over the past century, and signs of increasing rates of change have been observed in the last several decades. The consequences of climate change on the arctic freshwater system are varied and significant. Various authors report on receding glaciers, thinning and dwindling sea ice, recent record melt of the Greenland ice sheet, rising surface air temperatures, warming permafrost, treeline advance, and shrub encroachment on northern tundra landscapes [e.g., *Hinzman et al.*, 2005; *Serreze et al.*, 2000; *Overpeck et al.*, 1997; *Smith et al.*, 2005; *Sturm et al.*, 2003; *Stroeve et al.*, 2005]. Effects of climate change on the arctic freshwater system, and vice versa, are diverse and complex. While warming is relatively certain, climate variables such as precipitation, impacts on river discharge, and societal responses depend on many variables and on details of their interactions. Changes observed and expected on land, in the atmosphere, and the oceans are discussed according to the organization shown in Table 1. The potential impacts of these changes are also discussed but have received comparatively less study to date.

2. Land Changes

[3] In northern regions, hydrologic processes on land are primarily controlled by the presence or absence of permafrost. Hydrologic processes are also influenced by the

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Table 1. Observed Changes to the Arctic Freshwater Land, Atmospheric, and Ocean Systems[a]

Variable	Period of Record	Current Trend and Confidence
	Land Changes	
Permafrost Storage		
Active Layer Thickness –		
–Eurasia	1956–2000	+/confident
–North America	1990–2002	no trend/ confident
Soil Moisture	1960–2000	±/confident
Lakes/Wetlands		±/confident
River Q –		
–Eurasian	1936–2003	+/confident
–North American	1975–2000	No trend/ confident
–Hudson Bay	1964–2000	–/confident
	Atmospheric Changes	
Atmospheric Moisture Transport		+/confident
Atmospheric Storage		+/confident
Precipitation		
–over land N.A.		–/confident
–over land Eurasia		no trend/ confident
–over ocean		no trend/ uncertain
Precipitation – Evaporation		
–over land N.A.		+/confident
–over land Eurasia		uncertain
–over ocean		+/confident
	Ocean Changes	
Arctic Ocean		–/confident
North Atlantic/Nordic Sea		+/confident
Labrador Sea		+/confident
Sea ice		
–area	1978–pres.	–/very confident
–volume	1960s–pres.	–/confident
–first year	1978–pres.	+/very confident
Bering Strait	1999–pres	Decreasing between 1999 and 2001 Increasing between 2001 and 2004/ confident
Fram Strait outflow – liquid		
Fram Strait outflow		
–ice volume	1979–2002	+/confident
–ice area		
Canadian Arch. outflow		
–liquid		+/confident
–ice		
Hudson Strait		unknown

[a]Stocks are shown in bold and fluxes are shown in italics. A plus or minus sign indicates the direction of the trend. The confidence in the trend is referred to as confident, very confident, no trend, or uncertain. The period of record was based on the data available.

thickness of the active layer and the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in areal extent, the interaction between surface runoff and subpermafrost groundwater becomes more important [*Woo*, 1986]. The inability of soil moisture to infiltrate to deeper groundwater zones owing to ice-rich permafrost results in very wet surficial soils. In this section, we discuss changes observed in permafrost, soil moisture, lakes and wetlands, rivers, and snow cover.

2.1. Permafrost in the Arctic

[4] Arctic permafrost and active-layer conditions have undergone significant changes in recent decades. Change in permafrost temperature can be used as a sensitive indicator of the change in the long-term climate and surface energy balance [*Lachenbruch and Marshall*, 1986]. Very small changes in surface conditions can produce significant changes in permafrost temperatures [*Esch*, 1982; *Lachenbruch and Marshall*, 1986; *Osterkamp et al.*, 1994]. Monitoring of the permafrost thermal regime goes back as far as the 1940s in northern Alaska. Overall, permafrost surface temperature has increased 2 to 4°C over the last 50–100 a on the North Slope of Alaska [*Lachenbruch and Marshall*, 1986], although some sites showed little warming or even a cooling trend. Increasing air temperatures in the late 1880s and early 1900s in North America may have preceded warming of the permafrost. On the basis of data from Barrow, Alaska, changes in air temperature alone since the early 1920s could not explain the observed permafrost warming, implicating changes in snow cover or perhaps an earlier warming [*Zhang and Osterkamp*, 1993]. Recent measurements show that permafrost on the Alaskan arctic coastal plain and foothills warmed about 3°C since the late 1980s [*Osterkamp*, 2005]. Long-term monitoring of permafrost temperatures through deep boreholes in a north/south transect across the North Slope of Alaska reveals variable warming over the last 25 a. Permafrost temperatures along the Alaskan arctic coast increased about 2 to 3°C and about 0.5 to 1.5°C in the discontinuous permafrost regions south of the Brooks Mountain Range since the mid-1980s [*Osterkamp*, 2003, 2005].

[5] Permafrost temperature data collected since the mid-1980s from monitoring sites (at depths between 10 and 20 m) south of Norman Wells in the Mackenzie River valley indicate that a general warming of permafrost has occurred, with larger increases observed at the more northerly sites [*Smith et al.*, 2005]. No significant trend in permafrost temperatures was observed in the southern Mackenzie River valley, where permafrost is thin (less than 10 to 15 m thick) and warmer than −0.3°C [*Couture et al.*, 2003]. The absence of a trend is likely due to absorption of the latent heat required for phase change [e.g., *Riseborough*, 1990]. Similar results were reported for warm permafrost in the southern Yukon Territory [*Burn*, 2002; *Haeberli and Burn*, 2002]. In the central Mackenzie River valley, a temperature increase of up to 0.03°C a^{-1} was observed in permafrost with mean annual near-surface temperatures near −1°C and permafrost thickness of less than 50 m [*Smith et al.*, 2005]. *Kershaw* [2003] reports an increase in shallow (1.5 m) mean annual permafrost temperatures between 1990 and 2000 at some palsa sites to the west of the Mackenzie River valley, in the Macmillan/Caribou Pass region of the Northwest Territories. In the northern Mackenzie River basin, in colder (−7°C) and thicker permafrost, temperatures at a depth of 28 m increased by up to 0.1°C per annum in the 1990s [*Smith et al.*, 2005; *Couture et al.*, 2003]. Permafrost warming since the 1960s and 1970s in the Mackenzie Delta region has also been observed by *Burn* [2002]. This warming in the central and northern Mackenzie River region is consistent with the general increase in air temperature observed since the 1970s.

[6] Permafrost warming was also observed in the eastern Canadian high Arctic, but this appears to have mainly

occurred in the late 1990s. At the high arctic observatory at Alert, Nunavut, a warming of 0.15°C per annum (depth of 15 m) occurred between 1995 and 2001, and warming of about 0.06°C per annum (depth of about 30 m) occurred since 1996 [*Smith et al.*, 2003]. Shallow permafrost temperatures (2.5 m) increased by 1°C between 1994 and 2000 at a high arctic site at Lake Hazen, Ellesmere Island [*Broll et al.*, 2003]. Permafrost cooling was observed from the late 1980s to the early 1990s (depth of 5 m) at Iqaluit in the eastern Arctic. This cooling, however, was followed by warming of 0.4°C per annum between 1993 and 2000 [*Smith et al.*, 2005]. This trend is similar to that observed in Northern Quebec, where permafrost cooling (depth of 10 m) was observed between the mid-1980s and mid-1990s [*Allard et al.*, 1995], followed by warming beginning in 1996 [*Brown et al.*, 2000].

[7] The active layer is that portion of soil above permafrost that seasonally undergoes freezing and thawing. The active layer plays an important role in cold regions because most ecological, hydrological, biogeochemical, and pedogenic activity takes place within it [*Hinzman et al.*, 1991; *Kane et al.*, 1991]. Changes in active layer thickness are influenced by many factors, including surface temperature, physical properties of the surface cover and substrate, soil moisture, and duration and thickness of snow cover [*Brown et al.*, 2000; *Frauenfeld et al.*, 2004; *Zhang et al.*, 2005]. When other conditions remain constant, changes in active layer thickness can be expected to increase in response to climate warming, especially summer air temperature [*Zhang et al.*, 1997; *Brown et al.*, 2000]. Often, changes in active layer depth are masked by ground surface subsidence caused when ice-rich soils thaw at the bottom of the active layer [*Overduin and Kane*, 2006].

[8] Long-term monitoring of active layer thickness was conducted over the past several decades in Russia. By the early 1990s, about 25 stations, each containing 8–10 plots and 20–30 boreholes to depths of 10–15 m, measured ground temperatures [*Pavlov*, 1996]. More than 30 stations in the former Soviet Union, most started in the 1950s but a few as early as the 1930s, carried out measurements of soil temperature in permafrost. From 1956 to 1990, the active layer depth increased by about 20 cm [*Frauenfeld et al.*, 2004]. Changes in air temperature, thawing index, and snow depth were believed responsible for the increase in active layer thickness.

[9] The Circumpolar Active Layer Monitoring (CALM) program, developed in the 1990s, currently incorporates more than 100 sites worldwide [*Brown et al.*, 2000]. CALM is designed to observe the response of the active layer and near-surface permafrost to climate change. Results from northern high-latitude sites demonstrate substantial interannual and interdecadal fluctuations in active layer thickness. During the middle to late 1990s in Alaska and northwestern Canada, maximum and minimum thaw depth was observed in 1998 and in 2000, respectively, corresponding to the warmest and coolest summers. Evidence of increased active layer thickness, thaw subsidence, and thermokarst development were observed, indicating degradation of warmer permafrost [*Brown et al.*, 2000]. Intensive efforts to monitor active layer thickness throughout the circumpolar regions have not consistently found a thickening of the active layer. This may be due to melting of ice lenses at the base of the active layer, resulting in a widespread subsidence of the surface [*Overduin and Kane*, 2006]. Although thickening of the active layer may increase as the average annual air temperature increases, it may be difficult to detect in a thermokarsting, subsiding environment.

[10] Thermal degradation of ice-rich permafrost with coincident subsidence of the ground surface has recently resulted in extensive thermokarsting and creation of new water-filled surface depressions in northern Alaska (Figure 1) [*Jorgenson et al.*, 2001]. Analysis of aerial photography indicated that widespread ice wedge degradation had not occurred before 1980. Field observations and sampling also showed that ice wedge degradation has been relatively recent, as indicated by newly drowned vegetation. Despite the relatively cold average annual temperature of northern permafrost, thermokarst was widespread on a variety of terrain conditions, but most prevalent on ice-rich centers of old, drained lake basins and alluvial-marine terraces. Disturbance of the ground surface, which would have a similar effect on ice wedges, was not evident. As such the natural degradation was attributed to warmer weather of the recent decades [*Jorgenson et al.*, 2001].

2.2. Soil Moisture

[11] Soil moisture is the land surface hydrologic variable that most strongly affects land-atmosphere moisture and energy fluxes such as evaporative heat flux, and phase change in freezing or thawing of wet soil. Field measurements of soil moisture have been collected on the North Slope of Alaska, with emphasis on establishing macroscale and microscale topographic influences. Sites were installed in the foothill regions and on the coastal plain of the Kuparuk River basin. Preliminary results indicate that macrotopographic gradients greatly influence the importance of lateral versus vertical fluxes. Microtopographic differences affect the small-scale spatial differences in soil moisture but have less impact on flux direction. Soil moisture dynamics are controlled primarily by recent weather, topographic position, and presence or absence of permafrost; however, these dynamics are also markedly influenced by temperature and soil type [*Luthin and Guymon*, 1974].

[12] In relatively flat areas, where the ice-rich frozen layer is near the surface, soil moisture content is usually quite high. These areas have relatively high evapotranspiration and sensible heat transfer but low conductive heat transfer owing to the insulating properties of thick organic soils. As the active layer thickens throughout the summer, its capacity to store water increases, resulting in a time-varying basin response to storm events [*Bolton et al.*, 2000]. In some cases, the surficial soil may become drier when permafrost degrades and subsurface drainage improves. Thermokarst formation can lead to very high soil moisture levels in the bottom of the thermokarst structure and can cause drastic drying at the surface of the adjacent land.

2.3. Lakes and Wetlands

[13] Lake initiation, development, and disappearance are natural processes that occur in permafrost regions throughout the world [*Sellmann et al.*, 1975]. Recently, it has become apparent that the changes in lake surface area reported in many locations throughout the Arctic and Subarctic display consistent trends linked to changes in

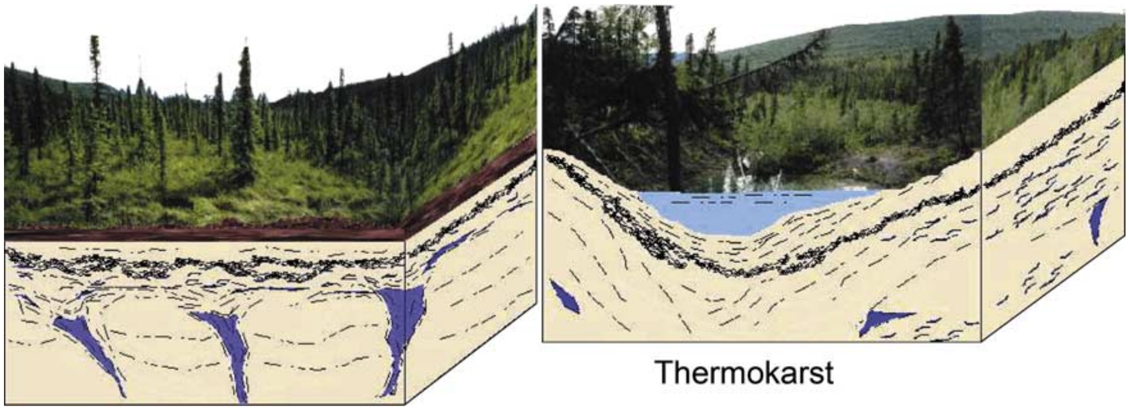

Figure 1. Thermokarst topography forms as ice-rich permafrost thaws and the ground surface subsides into the resulting voids. The important and dynamic processes involved in thermokarsting include thaw, ponding, surface and subsurface drainage, surface subsidence, and related erosion. These processes are capable of rapid and extensive modification of the landscape and preventing or controlling anthropogenic thermokarsting is a major challenge for northern development.

climate [*Hinzman et al.*, 2001]. Both increases and decreases in lake area have been documented. Such alterations can result from changes in local climatic processes but are also affected by consequences of thawing permafrost. Changes in lakes resulting from direct change in precipitation or evaporation are less evident because these subtle trends are within the large annual variation in weather. Surface water bodies may accumulate (increase in size) as permafrost begins to degrade and water accumulates, or lakes and ponds may decrease in size as deeper permafrost thaws, allowing internal drainage of the lake to groundwater. Lakes in permafrost regions may also drain catastrophically as surface channels develop when near-surface ice formations melt.

[14] In response to an imposed disturbance to permafrost, new surface depressions soon form ponds, which accelerate subsurface thaw through their lower albedo and additional heat advected into the pond through runoff. Such processes may tend to increase the number or size of lakes in areas of ice-rich, thick permafrost, such as on the Alaskan arctic coastal plain. In time, a talik may form below such ponds as the depth of water becomes greater than the amount that can refreeze during the winter. If the talik grows to a size that completely penetrates the underlying permafrost or connects to a subsurface layer that allows continued drainage, the pond may then begin to drain or become a discharge point of groundwater [*Kane and Slaughter*, 1972].

[15] A study near Council, Alaska investigated the physical factors that influence thermokarst pond formation [*Hinzman et al.*, 2001; *Yoshikawa and Hinzman*, 2003]. These studies demonstrated that in a warming climate, permafrost will degrade and ponds can drain if the lake water level is higher than the elevation of the groundwater table. Twenty-four small ponds on the Seward Peninsula were examined to determine if recent changes in climate had affected the dynamics of their development and degradation. Included in these studies were coring and thermal analyses, ground-penetrating radar surveys, and historical analyses through archival photographs and satellite imagery. Of the 24 ponds studied, 22 decreased in area between 1951 and 2000. The two that increased in size appeared to be controlled by open talik geometry and groundwater potential, while the remainder appeared to be perched over permafrost. Ground-penetrating radar studies near ponds that decreased in size showed taliks below them, allowing subsurface drainage to occur throughout the year. Analysis of meteorological data from Nome, Alaska did not show the marked temperature or precipitation differences likely to cause drying of the ponds. *Riordan et al.* [2004] suggest that losses in lake area observed in the Yukon Flats region of Alaska may also have been due to internal drainage or increased evaporation rates resulting from substantially rising temperatures over the last 30 a. During the same period, there was no observed trend in precipitation observed.

[16] *Smith et al.* [2005] described a widespread decline in lake abundance and area in the discontinuous, sporadic, and isolated permafrost zones across two large tracts of western Siberia, despite slight precipitation increases (Figure 2). That study also documented a net increase in lake area and the number of lakes exceeding 40 ha in size in the continuous permafrost zone. A recent study of change in lake area along the Kolyma River in northeastern Siberia [*Walter et al.*, 2006] suggested an increase of 14.7% on a 12,000 km^2 territory between 1974 and 2000, with the greatest increase occurring in lakes on ice-rich soils. These findings are counter to those of a lake change study conducted a year earlier on a smaller region (1700 km^2), which indicated a net decrease in lake area owing to the drainage of several large floodplain lakes located near major rivers (K. Walter, personal communication, 2007). Both analyses are consistent with the hypothesis that permafrost degradation and increased thermokarsting are the underlying mechanism of lake area change in this region of continuous permafrost.

[17] It is important to note that the changes in lakes described here are not occurring everywhere. Analyses of remotely sensed data on the North Slope of Alaska and the Old Crow Flats, two regions underlain by continuous permafrost, show no clear trend in pond/lake drying or expansion over the same time period. In these regions, some

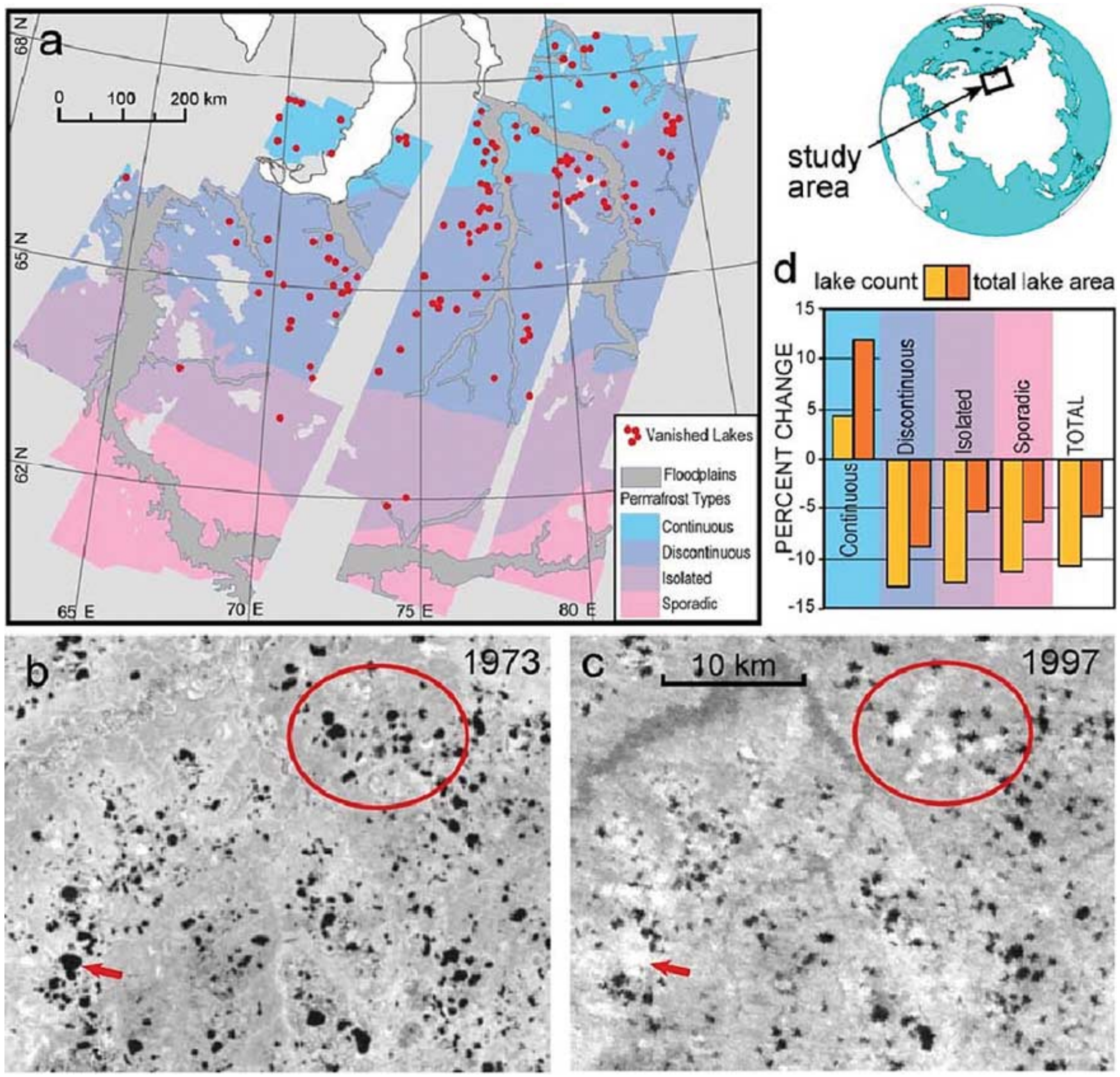

Figure 2. (a) Locations of Siberian lake inventories, permafrost distribution, and vanished lakes. (b) Total lake abundance and inundation area have declined since 1973, (c) including permanent drainage and revegetation of former lakebeds. (d) Interestingly, net increases in lake abundance and area have occurred in continuous permafrost, suggesting an initial but transitory increase in surface ponding [*Smith et al.*, 2005]. Reprinted with permission from AAAS.

ponds and lakes are decreasing in area while others are increasing (J. L. Labrecque et al., personal communication, 2007).

2.4. Arctic River Discharge

[18] River discharge is one of the best means of detecting large-scale changes in the terrestrial hydrologic cycle because it integrates precipitation, evapotranspiration, and changes in water storage over the entire watershed area upstream of the gauging station. This is certainly true in the Arctic, which contains some of the largest rivers (and watersheds) on Earth. Monitoring river discharge makes it possible to assess changes in the water budget over a huge part of the Arctic by monitoring a relatively small number of rivers. For example, each of the four largest rivers flowing into the Arctic Ocean, Yenisey, Lena, Ob', and Mackenzie, have watershed areas larger than the state of Alaska; the combined watershed area of these four rivers alone approaches 10 million km^2.

[19] A growing number of publications document substantial long-term changes in arctic river discharge. The longest records come from Russia, where monitoring typically began in the 1930s. Russia also contains the three largest arctic rivers and contributes the majority of the river water flowing into the Arctic Ocean. The emerging conclusion is that discharge from arctic rivers in Russia has increased over the period of record, with most of the increase occurring over the past few decades [*Savelieva et al.*, 2000; *Peterson et al.*, 2002; *Yang et al.*, 2002; *McClelland et al.*, 2004]. In an analysis of the combined discharge from the six largest Russian arctic rivers, *Peterson et al.* [2002] documented a 7% increase over 1936–1999. They showed that the trend in river discharge correlated with trends in both global surface air temperature and the North Atlantic Oscillation (NAO).

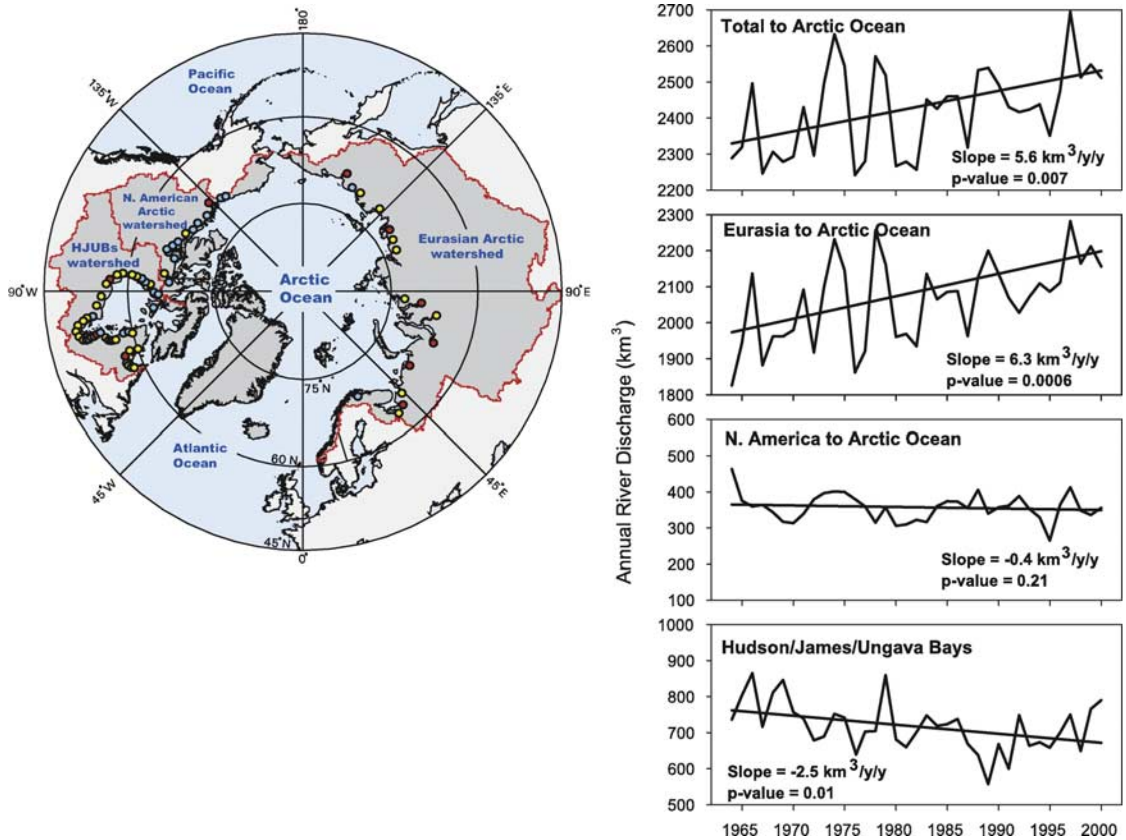

Figure 3. Map showing watersheds of the Arctic Ocean and Hudson, James, and Ungava Bays (HJUBs), with discharge trends for rivers draining these different watersheds shown in the accompanying graph. Colored dots on the map mark the mouths of the 72 rivers included in the analysis, with colors indicating discharge categories. Blue represents <6 km^3 a^{-1}, yellow represents 6 to <60 km^3 a^{-1}, and red represents 60–600 km^3 a^{-1}.

There has been lively discussion about the driver(s) of the observed increase in Eurasian arctic river discharge, with changes in precipitation including partitioning of rain and snow, permafrost thaw, and changes in fire regimes receiving the most attention [*Yang et al.*, 2002; *McClelland et al.*, 2004; *Berezovskaya et al.*, 2004; *Yang et al.*, 2004a, 2004b; *Wu et al.*, 2005].

[20] In contrast to the situation in Russia, discharge monitoring for downstream reaches of the major North American arctic rivers typically did not begin until the 1960s or 1970s. Recent analyses focusing on North American rivers showed small declines in discharge for rivers flowing into the Arctic Ocean [*Déry and Wood*, 2005], and substantial declines for those flowing into the Hudson Bay system (including James Bay, Ungava Bay, and Foxe Basin) [*Déry et al.*, 2005]. Detailed analyses of Alaskan rivers demonstrated increasing trends in discharge from glacial fed rivers and decreasing trends in nonglacial fed rivers [*Hinzman et al.*, 2005].

[21] Though these studies identified regional trends in arctic river discharge, differing periods of analysis and analytical methods have hampered comparison of river discharge trends for the Arctic as a whole. However, by restricting their analysis to the 1964–2000 period, when data availability was greatest throughout the pan-Arctic watershed, *McClelland et al.* [2006] documented that annual discharge into the Arctic Ocean changed by 5.6 km^3 a^{-1} (Figure 3). This increase was the net result of an annual increase 6.3 km^3 a^{-1} from Eurasia balanced by a small annual decline of -0.4 km^3 a^{-1} from North America. They also considered rivers flowing into Hudson Bay (including James and Ungava bays), where annual discharge decreased by 2.5 km^3 a^{-1}. Together the Arctic Ocean and the Hudson Bay system combined showed an increase in annual discharge of about 120 km^3 over the period 1964–2000 [*McClelland et al.*, 2006].

[22] In addition to long-term changes in annual discharge, there have been large changes in the seasonality of discharge. For example, several studies have reported substantial increases in winter and spring discharge for the major Eurasian arctic rivers [*Lammers et al.*, 2001; *Yang et al.*, 2002; *Serreze et al.*, 2002; *Ye et al.*, 2003; *McClelland et al.*, 2004; *Yang et al.*, 2004a; *Yang et al.*, 2004b]. The earlier studies [*Lammers et al.*, 2001; *Yang et al.*, 2002; *Serreze et al.*, 2003] tend to attribute these increases mainly to warming in winter, whereas later studies [*Ye et al.*, 2003; *McClelland et al.*, 2004; *Yang et al.*, 2004a, 2004b] tend to emphasize the confounding impact of hydroelectric dams, which shift discharge to the low-flow winter months to augment power generation. This does not mean that climate change does not impact the seasonality of discharge; it simply indicates that the story is complicated by regulation of dam discharge [*Georgievsky et al.*, 1996].

2.5. Arctic Snow Cover

[23] Seasonal snow cover is important in the global climate system owing to its effect on energy and moisture budgets. Snow cover is the largest single component of the cryosphere, with a mean maximum extent in the Northern Hemisphere of approximately 47 million km^2 or nearly 50% of the land surface area including the entire Arctic [*Robinson et al.*, 1993]. Seasonal snow cover is responsible for the largest seasonal and interannual differences in land surface albedo. Snow cover also significantly affects hydrologic processes, particularly in the high-latitude regions, owing to its direct impact on winter water storage and year-round runoff. Climate change significantly affects snow cover patterns over the high-latitude regions. Warming will likely result in higher precipitation, northward movement of the seasonal snow cover boundary, and a deeper snow cover overall in the Arctic [*Serreze et al.*, 2000; *Ye*, 2001a; *Ye et al.*, 1998]. Regional investigations documented large interannual and interdecadal variations and significantly increasing occurring trends in winter snow accumulation over the Siberian regions, with the most dramatic increases in the last 30 a, perhaps owing to increases in cold season precipitation [*Fallot et al.*, 1997; *Karen and Smith*, 2003; *Wang and Cho*, 1997; *Ye*, 2001b].

[24] Changes in timing, duration, thickness, and distribution of seasonal snow cover significantly impact surface runoff, groundwater recharge, and river streamflow. Interannual variation and changes in snow cover affect the seasonal and yearly runoff characteristics. Many studies have identified direct and positive correlations between basin snow water equivalent (SWE) and snowmelt runoff; that is, a correspondence of high (or low) snowmelt runoff with high (or low) snow cover at the end of winter season [*Kane et al.*, 2000; *Woo*, 1986]. Changes in seasonal streamflow patterns have been detected over the arctic rivers owing to changes in basin snow cover characteristics. For instance, studies show that over the recent decades snowmelt has started earlier in the northern regions, a trend associated with strong warming in winter and spring seasons [*U. S. Geological Survey*, 2000; *Yang et al.*, 2002; *Zhang et al.*, 2000]. These changes in the snowmelt and runoff patterns indicate a hydrologic regime shift toward earlier peak flow over the high latitudes. *Yang et al.* [2003, 2007] identified relationships between streamflow and basin snow cover extent and SWE during the spring melt period over the Ob, Yenisei, and Lena basins. More efforts are needed to better determine snow cover change and its impact on regional hydrology.

3. Atmospheric Changes

3.1. Changes in Atmospheric Moisture

[25] The atmospheric moisture budget can be separated into three major components: storage, horizontal (poleward) moisture transport, and net precipitation. Over land, relative changes in these parameters can be assessed with reasonable accuracy, as observations are available from rawinsondes and precipitation gauges. Long-term measurements of river runoff also aid in determining variability in regional net precipitation. An effort to explain observed increasing trends in river discharge by *McClelland et al.* [2004], however, could not definitively attribute a driver for the trends, but by process of elimination it appeared that changes in poleward moisture transport (PMT) were the most likely cause. This conclusion is supported by *Pavelsky and Smith* [2006], who compared several precipitation estimates with river discharge, but it is not supported by another recent comparative study [*Berezovskaya et al.*, 2004]. The lack of definitive agreement among precipitation data sets and the relationship to river discharge is a major obstacle to clarifying change mechanisms in the arctic terrestrial system. Over the Arctic Ocean an assessment of atmospheric moisture components is even less certain, as direct measurements of water vapor profiles, winds, and net precipitation are spatially and temporally sparse. A number of studies have attempted to estimate changes in atmospheric moisture; this section summarizes the most recent efforts. While the signs of change appear to be clear, especially in the past few decades, confidence in the trend magnitudes is less so.

3.2. Storage: Precipitable Water (PW)

[26] Atmospheric moisture storage refers to the amount of water vapor in the column of atmosphere. In the Arctic, where the lower atmosphere is frequently nearly saturated, this amount can be substantially affected by an increase in column temperature (and hence the saturation vapor pressure or vapor-containing capacity of the air). Thus as atmospheric temperature increases (or decreases), PW also typically increases (or decreases). PW also varies depending on the net horizontal transport of moisture into the column and by fluxes of moisture to or from the surface (condensation, sublimation, or evaporation).

[27] Trends in PW over the Arctic from 1982 to 1999 were calculated from output produced by the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis by *Wang and Key* [2005]. Over arctic land areas north of 60°N, they found no definitive change in the annual PW, which is consistent with similar calculations by *Serreze et al.* [2006], who used the reanalysis produced by the European Center for Medium-Range Weather Forecasting (ECMWF). Over the Arctic Ocean, however, *Wang and Key* [2005] found that PW decreased significantly in winter (0.001 cm a^{-1}) and increased during spring, summer, and fall (0.0014, 0.0030, and 0.0021 cm a^{-1}). These findings are consistent with observed changes in surface temperatures, and they support that saturation vapor pressure does control PW [*Serreze and Etringer*, 2003].

[28] Estimates of change in PW by season over the Arctic were also derived from satellite sounder retrievals. During 1979 to 2005, precipitable water increased by approximately 5 to 10% per decade during spring (MAM), with largest increases in the Beaufort/Chukchi Seas and also in the Western Arctic Ocean (Figure 4). In summer the PW increases are more uniform across the Arctic Ocean, except near the pole, with values of about 5% per decade. In winter and autumn, PW declined over much of northern Eurasia land areas, while increasing over most of the Arctic Ocean. These patterns closely resemble surface temperature changes retrieved from the Advanced Very-High Resolution Radiometer (AVHRR) by *Comiso* [2003] (updated to 2005 in the work of *Serreze and Francis* [2006]), further implicating the close link between the nearly saturated atmosphere and temperature. Because water vapor is a potent greenhouse gas,

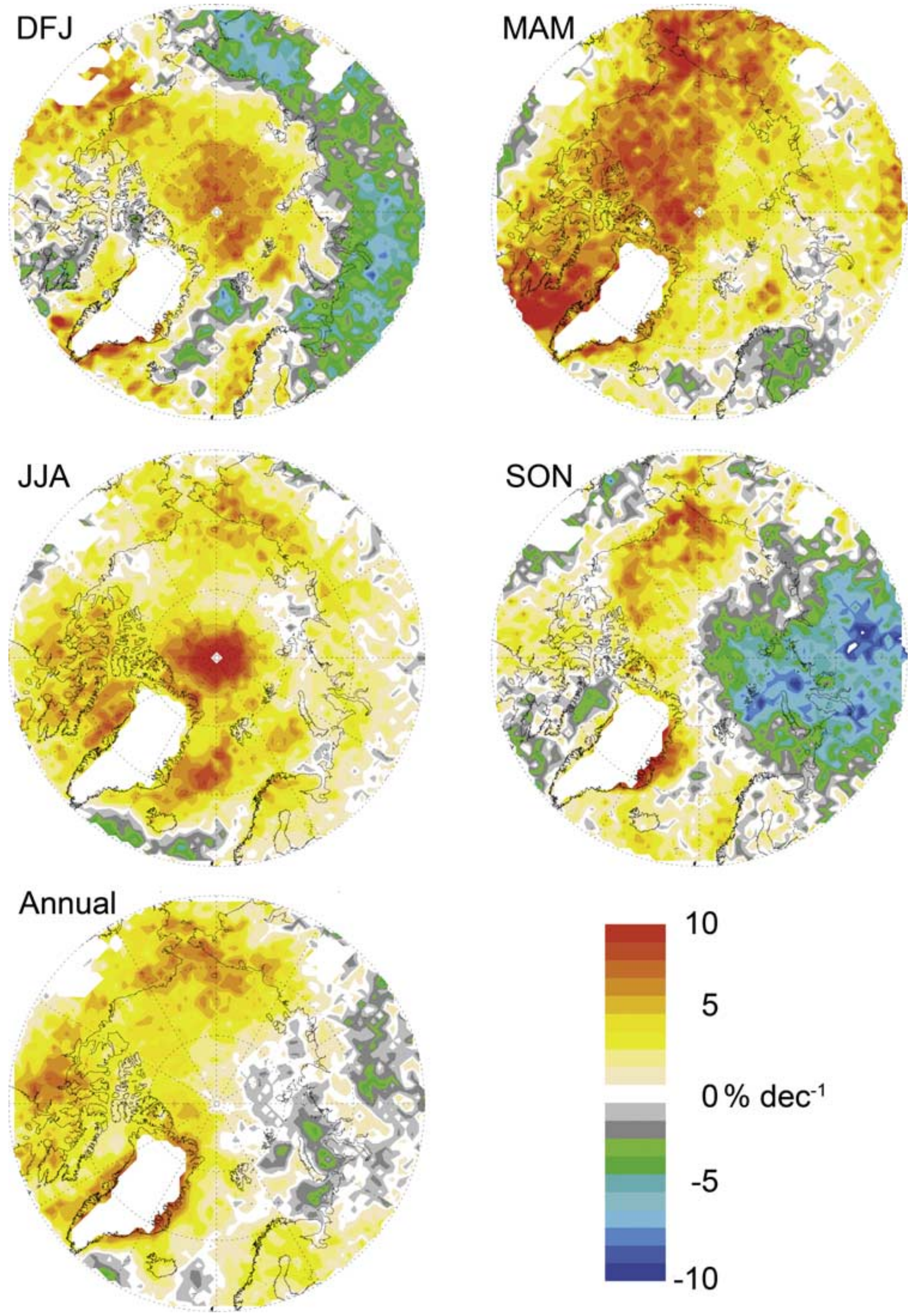

Figure 4. Seasonal trends in precipitable water derived from the TIROS Operation Vertical Sounder (TOVS) from 1979 to 2005. Trends are expressed as a change in the percent of the mean PW in each season per decade.

increases in PW directly contribute to increasing downwelling infrared radiation fluxes and therefore constitute a positive feedback on surface heating.

3.3. Poleward Moisture Transport (PMT)

[29] The contours of precipitable water, or total column water vapor content, are generally zonal in all seasons [e.g., *Groves and Francis*, 2002a]; the predominant source of moisture for the arctic region is horizontal transport poleward from lower latitudes. Changes in this flux may result from two effects: a change in the strength of the meridional moisture gradient, a change in the strength of the meridional wind, or a combination of both. Assessing changes in PMT requires accurate fields of precipitable water and winds. Once again, these variables are more reliable over land, where the rawinsonde network is adequate in most areas.

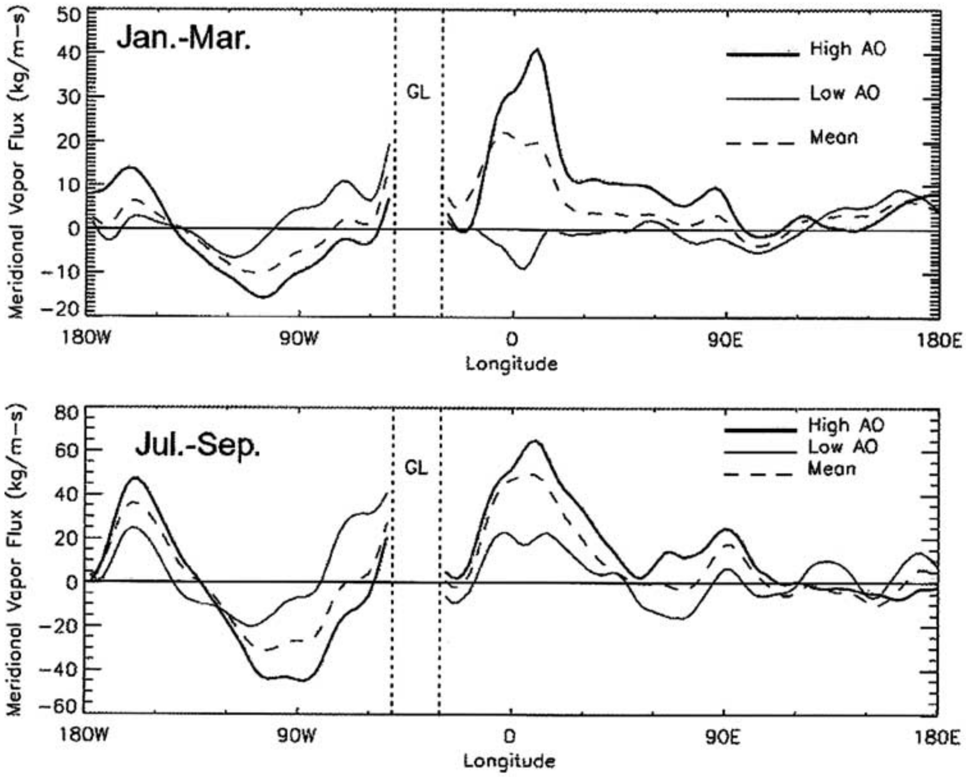

Figure 5. Composites of TOVS-derived poleward moisture transport across 70°N during 1980–1997 on high or positive AO-index days (thick solid) and low or negative AO-index days (thin solid) relative to the seasonal mean value (dashed) during (a) winter (January to March) and (b) summer (July to September). "GL" indicates the longitude band occupied by Greenland. Note differences in vertical scales [from *Groves and Francis*, 2002b].

[30] Calculation of PMT and its changes are based on analyzing three types of data: rawinsondes, satellite retrievals, and reanalyses, which assimilate rawinsondes and satellite radiances. All three types of analyses suggest that PMT is strongly tied to large-scale atmospheric circulation patterns such as the NAO, Arctic Oscillation/Northern Annular Mode (AO/NAM, hereafter NAM), and/or the Pacific Decadal Oscillation (PDO) [e.g., *Rogers et al.*, 2001; *Groves and Francis*, 2002b]. Because these large-scale indices have varied greatly in recent decades, it is difficult to separate natural variability from superimposed change, but the data seem to suggest that PMT has increased, particularly in Atlantic sector of the Arctic. Figure 5 (from *Groves and Francis* [2002b]) presents satellite-derived composites of the winter and summer differences in PMT across 70°N during periods of positive and negative AO index. Because there is more moisture in the atmosphere in summer months, changes in PMT during summer dominate the annual signal. East of Greenland the PMT is clearly larger when the AO is in a positive phase, but this relationship is somewhat counteracted by a sector west of Greenland with stronger southward fluxes. Overall, the PMT is larger when the AO is positive, which agrees with other findings, such as those of *Rogers et al.* [2001], based on reanalysis output. Consequently, because the NAO/AO has tended to be more positive in recent decades, observations suggest that the PMT has also increased. Over the Arctic Ocean, most of which lies north of 70°N, the evidence is not so clear. Satellite estimates suggest that PMT has increased in some areas and decreased in others, with no clear indication of an overall change averaged across the Arctic Ocean [*Groves and Francis*, 2002b].

3.4. Net Precipitation (P-E)

[31] Apparent increases in river runoff, precipitable water, and PMT would be expected to correspond with increased P-E, as well, but there are numerous difficulties in obtaining accurate measurements. These include snow redistribution by the wind, undercatchment by gauges in high wind conditions, and lack of regional representativeness of gauge locations owing to heterogeneous topography. Over the Arctic Ocean, problems in obtaining accurate P-E estimates are compounded by the almost complete lack of gauges, except for one or two Russian drifting ice stations that existed in the central Arctic between the mid-1950s and 1990. While topography is less of a factor here, redistribution and undercatchment issues are still significant.

[32] Having noted these issues, the available gauge data between 55° and 85°N suggest that precipitation has increased during winter and spring, while perhaps decreasing slightly during autumn [e.g., *Serreze et al.*, 2000]. These are estimates of precipitation, not P-E. The contribution by evaporation is expected to be largest in late spring through

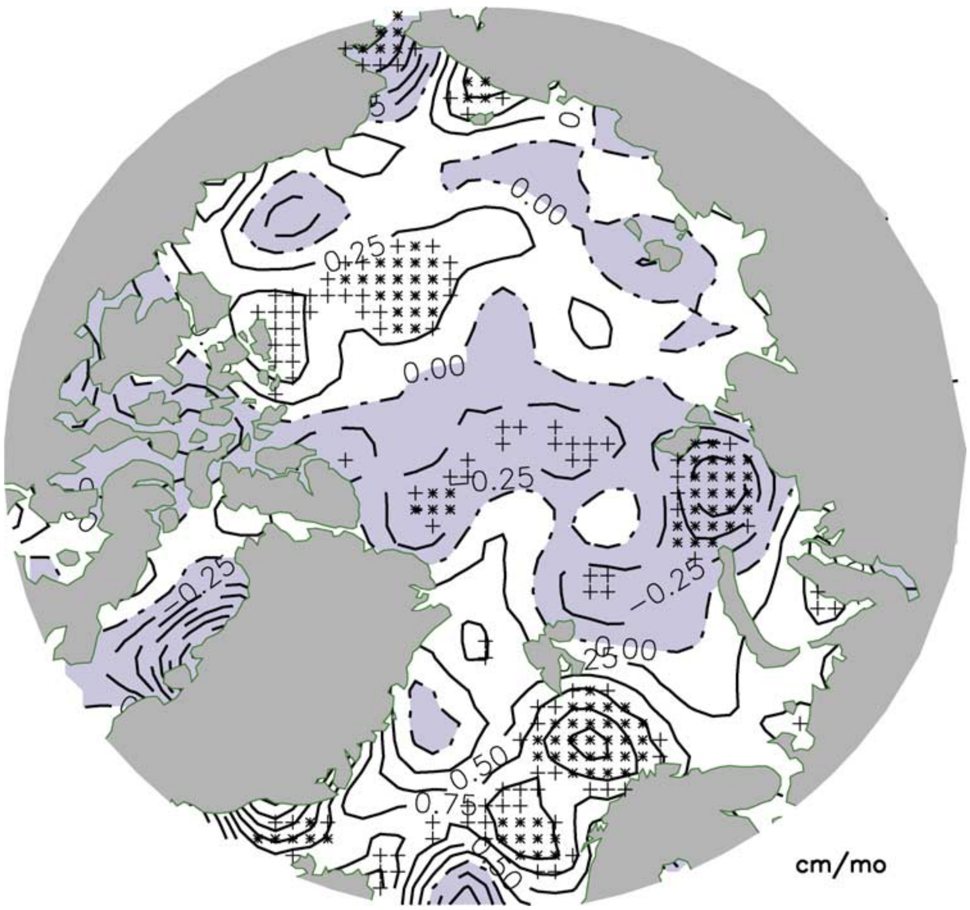

Figure 6. Decadal difference in annual mean satellite-derived P-E (1989 to 1998 minus 1980 to 1988) in cm/month. Shaded regions are negative differences, plusses indicate >90% confidence, and asterisks are >95% confidence [from *Groves and Francis*, 2002b].

early fall, when the surface is generally unfrozen and energy is available.

[33] Another approach to estimating P-E is often called the aerological approach, which involves calculating the convergence of moisture transport into an atmospheric column. If more moisture enters a column than leaves it, then the surplus minus the increased storage is designated "net precipitation." This approach has been successfully applied to rawinsonde data, reanalysis output, and satellite retrievals. Calculations of P-E from rawinsondes [e.g., *Walsh et al.*, 1994; *Serreze et al.*, 1995] and from reanalyses [e.g., *Cullather et al.*, 2000] exhibit no trends over the Arctic Basin north of 70°N between the early 1970s through the mid-1990s. Estimates from satellite retrievals over the Arctic Ocean (Figure 6) suggest that from the 1980s to the 1990s, P-E increased in much of the GIN Seas area, the Bering Strait, and the northern Beaufort Sea but decreased in the Kara Sea and near the North Pole [*Groves and Francis*, 2002b]. This pattern reflects the predominantly positive phase of the NAO/AO in the early 1990s, which would be expected to increase moisture transport into the North Atlantic [e.g., *Dickson et al.*, 2000]. Estimates from the European Centre for Medium-Range Weather Forecast (ECMWF), a reanalysis based on 40 a of data, suggested that P-E has increased slightly over the last half-century [*Peterson et al.*, 2006].

4. Ocean Changes

[34] The Arctic Ocean is a giant estuary, with an upper 200 to 400 m of low salinity ($29 < S < 34$) water fed by river discharge, precipitation, relatively fresh inflow from the North Pacific Ocean via the Bering Strait, and deeper, warmer, more saline ($S \sim 35$) waters maintained by Atlantic water inflow through the Fram Strait and the Barents Sea. Residence time for freshwater in the Arctic is approximately 10 a [*Östlund*, 1982; *Schlosser et al.*, 1990; *Ekwurzel et al.*, 2001], with both liquid and solid (ice) components exported southward to the North Atlantic. The strong halocline formed by the fresh upper layer insulates Arctic Ocean ice cover from heat imported by inflowing Atlantic waters, which could inhibit ice formation and accelerate melting. Heat flux associated with Atlantic inflow thus can play an important role in driving changes to the arctic heat balance and freshwater system.

[35] Freshwater exchange between the Arctic and North Atlantic provide a critical mechanism through which arctic variability and global climate interact. Arctic freshwater discharges through the Canadian Arctic Archipelago/Davis Strait, Fram Strait and the Barents Sea into deep water

formation regions to the west (Labrador Sea) and east (Greenland/Irminger Seas) of Greenland. Freshwater inflow contributes a buoyant surface layer that can act as a barrier, inhibiting convective overturning and deepwater formation and affecting the meridional overturning circulation (MOC). Recent assessments of decadal-scale variability in North Atlantic freshwater content [*Curry et al.*, 2003; *Curry and Mauritzen*, 2005] reveal systematic freshening in the receiving basins west and east of Greenland, while *Zweng and Münchow* [2006] use archived data to identify freshening along the western side of Baffin Bay, extending southward through Davis Strait and into the Labrador Sea. These studies implicate increased arctic freshwater outflow as the dominant contributor to growing freshwater inventories in these important MOC control regions, though estimated freshening falls below the levels required for MOC shutdown in modeling studies [*Rahmstorf et al.*, 2005].

[36] Ocean freshwater is often measured relative to a reference salinity, usually taken as a mean arctic salinity of $S_{\mathrm{ref}} = 34.8$ [*Aagaard and Carmack*, 1989]. Freshwater content may be defined as:

$$FWC = \int_{200\,\mathrm{m}}^{0\,\mathrm{m}} \frac{\left(S_{ref} - S(z)\right)}{S_{ref}} dz. \qquad (1)$$

and in terms of units of height. Averaged over the entire Arctic Ocean, the upper 200 m contains roughly three quarters of the total freshwater content. Throughout this section, unless stated otherwise, salinity is measured on the Practical Salinity Scale and is quoted with no unit.

4.1. Arctic Ocean Changes

[37] *Aagaard and Carmack* [1989] and, more recently, *Serreze et al.* [2006] summarize the literature on the freshwater storage of the Arctic seas. However, long-term multiyear trends in ocean freshwater content are not straightforward to calculate, owing to a lack of accessible data. The longest reliable time series are regional and come from the Russian arctic shelves, where summer data have been collected since the early decades of the 20th century, largely as a result of ship-based coastal community supply operations. *Steele and Ermold* [2004] used these data to determine that ocean salinities generally increased over 1930–1965 and then decreased over 1965–1995, in anticorrelation with the NAO/AO. They proposed that the eastern shelves (East Siberian Sea and Laptev Sea) are controlled by wind forcing; that is, high/low NAO/AO index implies more westerly/southerly winds which force freshwater along/across the shelf. For the Kara Sea, this mechanism may be enhanced by increased/decreased atmospheric moisture flux convergence during high/low NAO/AO index conditions, which increases both local P-E and river discharge. For the White Sea, salinity changes seem controlled mostly by river discharge trends [*Peterson et al.*, 2002].

[38] In the central Arctic Ocean, large-scale salinity observations were taken mostly by Russian scientists, starting about 1950 and using both drifting ice camps and springtime aerial surveys. These data are publicly available only as decadal average three-dimensional salinity (and temperature) fields [*Arctic Climatology Project*, 1997, 1998], from which it is difficult to determine a trend [*Häkkinen and Proshutinsky*, 2004], or as annual averages over specified Arctic Ocean subregions and depth ranges [*Swift et al.*, 2005]. Trends are also difficult to determine using numerical ocean models, because of model and forcing biases that frequently lead to unrealistic freshwater drift away from observations [*Steele et al.*, 2001a, 2001b]. Recently, *Steele and Ermold* [2007] used a version of the public Russian data that consists of annual (not decadal) mean fields of dynamic (or "steric") height from 1950 to 1990. Dynamic height is a vertical integral (in this case, over the upper 200 m) of the density departure from a reference value. Since density is so strongly controlled by salinity at the cold temperatures of the Arctic, this is essentially a measure of freshwater content. *Steele and Ermold* [2007] found a generally decreasing trend in dynamic height over the Arctic Ocean, equivalent to a loss of freshwater content of about 30 cm over 4 decades. That is, in spite of the gradually increasing freshwater sources from river discharge, P-E, and reduced net sea ice growth and changes in the Bering Strait inflow, the Arctic Ocean lost freshwater. How could this happen? The answer seems to be in accelerated export to the North Atlantic Ocean, which in fact did freshen over this time period [see also *Curry and Mauritzen*, 2005]. *Steele and Ermold* [2007] further noted that this export occurs mostly as discrete events, such as the Great Salinity Anomaly of the late 1960s through 1970s [*Dickson et al.*, 1988]. They found that, possibly forced by wind changes associated with the NAO/AO index, sea ice first quickly flushes out of the Arctic Ocean and follows the East Greenland Current into the subpolar gyre. If these winds are sustained, the upper few hundred meters of freshwater then slowly drain southward over the next roughly10 a, a phenomenon observed as a dynamic height and freshwater content drop in the Arctic Ocean and a gain in the North Atlantic Ocean [see also *Haak et al.*, 2003; *Karcher et al.*, 2005]. However, during the 1970s the North Atlantic gain was larger than the Arctic drop, implying another, as yet undetermined, source. Recently, *Scharroo et al.* [2006] used satellite altimeter data to determine that Arctic Ocean sea level continued to fall in the late 1990s and early 2000s, strongly suggesting a continued drop in freshwater content.

[39] Over perhaps the next few decades, researchers expect an accelerated hydrologic cycle may bring ever-increasing freshwater sources to the Arctic Ocean, resulting in a freshening ocean. However, the exact opposite seems to have occurred over the last half of the 20th century; that is, a change in wind forcing seems to have accelerated freshwater export more than the slowly increasing freshwater sources.

4.2. Sea Ice

[40] As sea ice grows from the ocean waters it rejects salt, resulting in a relatively fresh ice cover with a salinity of approximately 4. Thus the ice growth-melt annual cycle represents a redistribution of freshwater between these two (ice and ocean) reservoirs. The presence of the perennial arctic sea ice amounts to a considerable freshwater storage of approximately 10,000 km^3 [*Serreze et al.*, 2006], assuming a reference salinity of 34.8. Sea ice also moves in response to winds and ocean currents, resulting in a spatial redistribution of freshwater largely from the Arctic to the northern North Atlantic, where a net ice melt is present.

[41] Observations indicate that changes in both the sea ice storage of freshwater and the flux of freshwater out of the Arctic via ice transport are underway. Sea ice concentration has been well-observed from satellites starting in 1979. Over this observed record, from 1979 to 2005, the summertime ice extent has decreased at about 8% per decade [*Cavalieri et al.*, 2003; *Stroeve et al.*, 2005], with extreme minimum ice years observed in the 2000s [*Serreze et al.*, 2003b; *Stroeve et al.*, 2005]. This decreases perennial ice cover, and observations confirm a sharp reduction in the multiyear ice area [*Johannessen et al.*, 1999; *Comiso*, 2002]. These summertime trends are particularly large in the Eurasian Arctic [*Parkinson et al.*, 1999]. Smaller but still decreasing trends are also seen in the winter ice cover. These have been particularly noteworthy in recent years, with all months of the winter and spring of 2004–2005 (December to March) being well below normal [*Meier et al.*, 2005]. Accounting for these recent years, the downward trend in winter ice extent has approached 3% per decade since 1979.

[42] In concert with the changing ice extent, the sea ice thickness, which directly relates to the freshwater ice storage, has shown reductions since the 1960s [*Rothrock et al.*, 1999; *Wadhams and Davis*, 2000]. These ice volume changes are dramatic, particularly for the 1990s. Using upward looking sonar measurements, *Rothrock et al.* [1999] estimated a decrease of approximately 1.3 m over much of the deep water regions of the Arctic Basin, from a mean of 3.1 m in 1958–1976 to 1.8 m in 1993–1997. This is consistent with the 43% decrease in ice thickness *Wadhams and Davis* [2000] found for a transect between Fram Strait and the North Pole from 1976 to 1996. Some studies raised concerns about the sparse spatial and temporal sampling of these measurements [e.g., *Holloway and Sou*, 2002] and suggested that this meagerness may have lead to overestimation of the ice volume changes. However, *Rothrock et al.* [2003] found "compelling agreement" between sea ice model simulations and observations for the ice thickness decline from the late 1980s through 1997, suggesting that a decreasing trend, although perhaps not the magnitude of that trend, was robust over that time period.

[43] Ice transport out of the Arctic is related to both ice thickness and velocity; thus changes in the arctic ice thickness can directly modify this freshwater flux. However, changes in the ice motion are also present, and these complicate the picture. The primary ice flux from the Arctic is through Fram Strait. This ice then melts within the northern North Atlantic and modifies the surface ocean conditions there. This process can have important implications for the ocean deep water formation that occurs within the north Atlantic. The ice area transport through Fram Strait has been estimated from satellite-derived ice velocity measurements and ice concentration measurements [*Kwok and Rothrock*, 1999; *Kwok et al.*, 2004]. A strong relationship exists between the winter ice area export and the NAO index, which describes a dominant mode of atmospheric variability [*Kwok et al.*, 2004]. This highlights the importance of changing atmospheric circulation for the ice transport. Additionally, from 1978 to 2002 there was an upward trend in the winter ice area export of 3040 $km^2\ a^{-1}$ (from an average of 754,000 km^2). Upward looking sonar observations of ice draft across Fram Strait combined with estimates of ice velocity provide estimates of ice volume flux [*Vinje*, 2001; *Kwok et al.*, 2004]. These calculations involve considerably more uncertainty than the ice area flux estimates owing to inadequate sampling of the across-strait ice draft [*Kwok et al.*, 2004]. Using these methods, *Vinje* [2001] obtained a time series of ice volume transport from 1950 to 2000. This series exhibits high interannual variability, no trend in the wind-induced ice efflux, and shows little correlation with the NAO winter index. For shorter periods (1978–1997) a positive correlation with the NAO is present (this is consistent with *Kwok and Rothrock* [1999]).

4.3. Boundary Flux Changes

[44] The semienclosed Arctic Ocean communicates with lower-latitude basins through a small number of constricted passages. Relatively fresh Pacific waters enter through the 85 km wide, 50 m deep Bering Strait, which represents the only oceanic link between the Pacific and Arctic Oceans. The Bering Strait supports an annual mean northward flow, believed to be driven by a north-south pressure gradient and opposed by generally southward local winds [*Coachman and Aagaard*, 1966, 1981; *Woodgate et al.*, 2005a]. The largest freshwater exports occur through the 500 km wide, 2500 m deep Fram Strait and the complex series of narrow (5–40 km), shallow (100–400 m) straits that comprise the Canadian Arctic Archipelago, with a much smaller component exiting the broad opening to the Barents Sea (see Figure 8) [*Aagaard and Carmack*, 1989; *Serreze et al.*, 2006]. Further south, at the northern end of the Labrador Sea, Davis Strait integrates freshwater outflow from the majority of the Canadian Arctic Archipelago/Baffin Bay system, and Hudson Strait collects the discharge from narrow, shallow Canadian Arctic Archipelago passage (Fury and Hecla Strait) and the Foxe Basin/Hudson Bay system. Atlantic water enters the Arctic from Fram Strait and through the Barents Sea opening. Atlantic water also penetrates northward into Baffin Bay but is not believed to traverse the shallow straits of the Canadian Arctic Archipelago to enter the Arctic Ocean.

[45] Attempts to quantify fluxes through the critical arctic straits face severe challenges that limit the availability of data suitable for assessing decadal-scale change. All the critical straits are dynamically wide (strait width larger than internal deformation radius) and thus admit small-scale recirculation that must be resolved to accurately quantify fluxes. Because conventional oceanographic measurement approaches cannot generally achieve sustained kilometer-scale horizontal resolution, most existing records severely alias scales at 10 km and smaller, introducing large uncertainties to flux and transport calculations. Proximity to the magnetic pole complicates quantifying flow direction, which recent technologies [e.g., *Prinsenberg and Hamilton*, 2005; *Melling*, 2004] have worked to overcome. Iceberg keels, along with ridging associated with seasonal and permanent ice cover, threaten instruments moored within 300 m of the sea surface. This limits the availability of salinity measurements within the upper water column, where most freshwater resides. Most existing flux estimates proceed, at least to some degree, by assuming spatially and temporally invariant salinity fields. Although paucity of measurements forces these assumptions, they are acknowledged to be dubious [e.g., *Münchow et al.*, 2006; *Prinsenberg and Hamilton*, 2005], and they introduce large uncertainties

to the results. Moreover, freshwater transport within the ice must also be quantified, requiring measurements of both ice motion and draft spectrum. Finally, remoteness, harsh operating conditions, and, in the case of Bering Strait, political concerns have limited the scope and duration of measurement efforts in critical Arctic straits. Thus unlike some of the other arctic freshwater system components, only limited information exists for assessing flux variability and change at many critical gateways. Rather, the oceanographic community still wrestles with the difficulties of collecting measurements with the appropriate resolution, duration, and certainty to detect change [cf. *Melling*, 2000]. Significant measurement efforts, including technological developments tailored to address many of these challenges, are currently directed at all of the straits discussed below.

4.3.1. Bering Strait

[46] Although older flux estimates are higher (e.g., ~1.2 Sv from the 1950s [see *Mosby*, 1962]), estimates of the Bering Strait transport in recent decades set the flow at ~0.8 Sv [*Coachman and Aagaard*, 1981; *Roach et al.*, 1995; *Woodgate et al.*, 2005a, 2005b]. In a synthesis of arctic freshwater fluxes, *Aagaard and Carmack* [1989] used this transport estimate, and an assumed salinity of 32.5, to obtain a Bering Strait freshwater flux estimate of ~1670 km^3 a^{-1} (relative to 34.8, an estimate of the Arctic Ocean mean salinity). A more recent effort [*Woodgate and Aagaard*, 2005] attempts to account for vertical structure and the seasonal presence of the Alaskan Coastal Current (ACC), estimating freshwater flux of ~2500 ± 300 km^3 a^{-1}, including estimated contributions from the ACC (~220–450 km^3 a^{-1}), water column stratification of the central Bering Strait (~350 km^3 a^{-1}), and ice transport (~100 km^3 a^{-1}).

[47] Drawing on moored measurements collected in the Bering Strait between 1990 and present day, *Woodgate et al.* [2006] attempted to quantify interannual variability in freshwater flux through the strait. Between 1998 (the start of the continuous data record) and present day, freshwater fluxes peak in 1998 (~2000 km^3 a^{-1}), decrease to 2001 (~1300 km^3 a^{-1}), then increase again until the last year of data currently available, 2004 (~2100 km^3 a^{-1}). Each estimate has errors of ~200 km^3 a^{-1}, and the effects of stratification and the ACC (likely ~800–1000 km^3 a^{-1}) are not included in these estimates. *Woodgate et al.* [2006] noted that although there is some modest freshening, about 80% of the freshwater increase between 2001 and 2004 is attributable to an increase in northward volume transport (from ~0.7 Sv to ~1.0 Sv) over the same time period, which, in turn, is likely driven by a weakening of the southward winds. The increase in freshwater flux between 2001 and 2004 is about 800 km^3 a^{-1}, roughly one quarter of annual mean arctic river runoff. These figures suggest that local wind effects in the Bering Strait may have a sizeable influence on Arctic Ocean freshwater fluxes.

4.3.2. Fram Strait

[48] Fram Strait discharges arctic freshwater (liquid and ice in nearly equal proportion) into the deepwater formation regions of the Greenland and Irminger seas and, through its contribution to the West Greenland Current, to the eastern margin of the Labrador Sea. Roughly 10% of the total Arctic Ocean sea ice mass exits Fram Strait each year [*Kwok et al.*, 2004]. Fram Strait also offers the only deep passage between the Arctic and North Atlantic, through which warm, saline Atlantic waters pass northward. This represents a large heat flux into the Arctic that impacts stratification, circulation, ice formation and melt, and atmospheric variability [*Schauer et al.*, 2004].

[49] Fram Strait fluxes can be separated into southward flowing arctic waters, for which the focus is freshwater transport, northward flowing Atlantic waters (the West Spitsbergen Current), where heat content entering the Arctic is the primary focus and southward ice export. Early Fram Strait volume and liquid freshwater transport estimates were based on sparse moored arrays and limited hydrographic data. Arctic liquid freshwater southward export estimates range from 26 to 76 mSv [*Aagaard and Carmack*, 1989; *Meredith et al.*, 2001], West Spitsbergen Current volume transport estimates range from 5.6 to 7.1 Sv [*Aagaard and Greisman*, 1975; *Hanzlick*, 1983], and an early ice freshwater flux estimate is 90 mSv [*Vinje and Finnekåsa*, 1986]. More recently, a 3-a record collected by a moored array has yielded new volume flux estimates of 9–10 Sv northward (West Spitsbergen Current) and 12–13 Sv southward [*Fahrbach et al.*, 2001; *Schauer et al.*, 2004]. Though the data do not support investigation of decadal-scale change, the record does reveal northward heat transport increasing from 28 ± 5 to roughly 45 ± 5 TW, owing to both stronger northward flow and higher temperatures. Though this signal spans only the 3-a record, *Schauer et al.* [2004] pointed out that an increase of similar magnitude would provide enough additional heat to explain the warming in the Eurasian Arctic observed in the early 1990s. Using data from ice buoy drifts, Synthetic Aperture Radar images, and upward looking sonar measurements, *Vinje* [2001] assembled a 50 a (1950–2000) Fram Strait ice flux record. Although this time series suggests that large interannual variations in ice flux are typical, it revealed no significant long-term trends. In contrast, *Kwok et al.* [2004] employed 9 a of ice thickness data collected using upward looking sonar data to demonstrate a statistically significant decrease in mean ice thickness over Fram Strait; this study also related changes at interannual timescales to changes in the magnitude and pattern of ice circulation in the Arctic Ocean.

4.3.3. Canadian Arctic Archipelago

[50] Total volume fluxes through the three main passages of the Canadian Archipelago have been variously estimated through both models and observations to be 0.7–2.8 Sv [*Fissel et al.*, 1988; *Muench*, 1971; *Prinsenberg and Hamilton*, 2005; *Rudels*, 1986; *Steele et al.*, 1996]. Assessments of the arctic freshwater budget [*Aagaard and Carmack*, 1989; *Dickson et al.*, 2007; *Serreze et al. 2006*] suggest that freshwater flows associated with these volume fluxes (29–110 mSv) rival or exceed that of sea-ice (70–92 mSv) and freshened seawater (20–95 mSv) passing through Fram Strait. Owing to data limitations, differences in the literature between estimates for fluxes through the Canadian Archipelago through at least 2006 should be understood as the result of uncertainties in approach rather than true changes over time.

[51] While existing observational times series are insufficient to establish whether there are any trends in fluxes through the passages, clear signals of change can be gleaned from regional hydrography. Through examination of historical temperature and salinity data for Baffin Bay, researchers have shown that there has been a significant warming of

waters at and below the Davis Strait sill depth and a freshening of waters above 250 m [*Zweng and Münchow*, 2006] over the 1916–2003 period. The heat appears to originate from warming inflow in eastern Davis Strait, while the freshening derives from the arctic sources flowing through the Archipelago passages. Limited tracer hydrographic sections of the passages suggested higher summer fluxes in 1997, with respect to 1977 and 2003 (K. Falkner et al., Interannual variability of dissolved nutrients in the Canadian Archepelego and Baffin Bay with implications for freshwater flux, submitted to *Journal of Geophysical Research*, 2007). This may be related to atmospheric pressure changes, as the AO index was strongly positive in 1997, negative in 1977, and near neutral in 2003. Moored time series measurements currently underway should help reveal the nature of atmospheric and sea level connections to flux change.

4.3.4. Davis Strait

[52] Arctic waters exiting Davis Strait have undergone numerous transformations during transit through the Canadian Arctic Archipelago and Baffin Bay systems, including terrestrial inputs from Baffin Island and Greenland, an annual average atmospheric heat loss of roughly 70 Wm^{-2}, brine and freshwater inputs from sea ice production and melting, and mixing with the underlying warmer, more saline Atlantic waters [*Rudels*, 1986]. Changes within these elements of the arctic freshwater system manifest as variability in this integrated outflow. Because Davis Strait is the dominant freshwater source for the Labrador Sea [*Lazier and Wright*, 1993; *Loder et al.*, 1998], these changes will also be observed as water mass and circulation shifts over the shelf and central basin and, potentially, though broader changes over the North Atlantic.

[53] Existing Davis Strait volume and freshwater flux estimates stem from a 3-a time series (1987–1990) collected by a sparse moored array [*Ross*, 1992] that neglected the West Greenland Shelf and the upper 150 m of the water column. Excluding highly extrapolative flux estimates for the West Greenland Shelf, annual mean southward volume (freshwater) flux estimates range from 3.4 Sv (129 mSv) [*Cuny et al.*, 2005] to 2.6 Sv (99 mSv) [*Tang et al.*, 2004]; the differences are driven by the assumptions used to interpolate between mooring sites and to extrapolate from the 150-m depth to the sea surface. *Ingram and Prinsenberg* [1998] estimate Davis Strait ice freshwater flux at roughly 35 mSv. Although short record length and large uncertainties preclude discussion of decadal-scale change based on the moored time series, other observations offer limited evidence of change. *Houghton and Visbeck* [2002] used historical salinity data (1948–1997) to show significant decadal-scale salinity variability on the eastern side of Davis Strait and an overall freshening trend from 1948 into the early 1990s. In contrast to recent findings of reduced ice cover over much of the Arctic, *Stern and Heidi-Jørgensen* [2003] found an increasing trend in maximum ice extent across the Davis Strait region using satellite passive microwave data (1979–2001) and digitized Danish Meteorological Service ice charts (1953–1981). Last, comparison of recent hydrographic sections and moored observations with a 1928–2004 climatology suggest that warm, relatively saline Atlantic waters occupied an anomalously large portion of the Strait in 2004–2006 (C. Lee and B. Petrie, personal communication, 2007); these findings are consistent with the results of *Zweng and Munchow* [2005] that demonstrate warming in the upper layers of Baffin Bay.

4.3.5. Hudson Strait

[54] The Hudson Bay System (HBS) is a large, Arctic inland sea characterized by an unusually large freshwater input from rivers (roughly 10^3 km^3 a^{-1} [*Déry et al.*, 2005]) and, to a lesser extent, from the Arctic Ocean via Fury and Hecla Strait. Its principal opening is Hudson Strait, connecting it to the Labrador Sea. The circulation in the Strait is toward the HBS, along Baffin Island, and toward the Labrador Sea along the Quebec coast [*Drinkwater*, 1988; *Ingram and Prinsenberg*, 1998]. On the basis of on the freshwater input into HBS from rivers, precipitation minus evaporation, and the flow through Fury and Hecla Strait, the net freshwater transport out of Hudson Strait (toward the Labrador Sea) is estimated to be 38 mSv (relative to a salinity of 34.8 [*Straneo and Saucier*, 2007a]). This makes Hudson Strait the third largest contributor of freshwater export from the Arctic region to the North Atlantic, after Fram and Davis Straits [*Dickson et al.*, 2007].

[55] A recent year-round monitoring program of the outflow from Hudson Strait into the Labrador Sea (along the Quebec coast) found that the current along the southern edge of the strait transports approximately 80 mSv of freshwater [*Straneo and Saucier*, 2007b]. This suggests that approximately one third of the freshwater outflow from Davis Strait flows into Hudson Strait, instead of flowing directly into the Labrador Sea. Once in the Labrador Sea, the Hudson Strait outflow combines with that from Davis Strait in the Labrador Current, a freshwater-laden current flowing close to the region of dense water formation in the Labrador Sea, which is thought to play a significant role in modulating its occurrence and variability [*Lazier et al.*, 2002].

[56] Hydrographic data collected in the HBS is scarce; there are no records to show whether the sea water properties have changed or how. Nonetheless, there are strong indications that the system has undergone rapid change. The river runoff into HBS has decreased by 100 km^3 from 1964 to 2000 [*Déry et al.*, 2005], likely owing to shifts in precipitation [*Déry and Wood*, 2005]. This decrease balances approximately 40% of the observed increasing river discharge into the Arctic Ocean [*McClelland et al.*, 2006]. Similarly and potentially related, the extent of the sea-ice cover has also decreased over the last 2 decades [*Laine*, 2004]. All these changes are presumably associated with changes in the local freshwater storage and export, but the limited data available makes it difficult to assess them. A number of models indicate that warming owing to an increase in greenhouse gases will lead to an ice-free HBS [*Gagnon and Gough*, 2005; *Saucier and Dionne*, 1998; *Boer et al.*, 2000].

5. Impacts of Changes to the Arctic Freshwater Cycle

5.1. Impacts of Changes on Land

5.1.1. Permafrost and Soil Moisture Conditions

[57] Permafrost degradation and hydrologic change will cause marked changes to surface soil moisture levels and strongly influence wildfire behavior [*Hinzman et al.*, 2006].

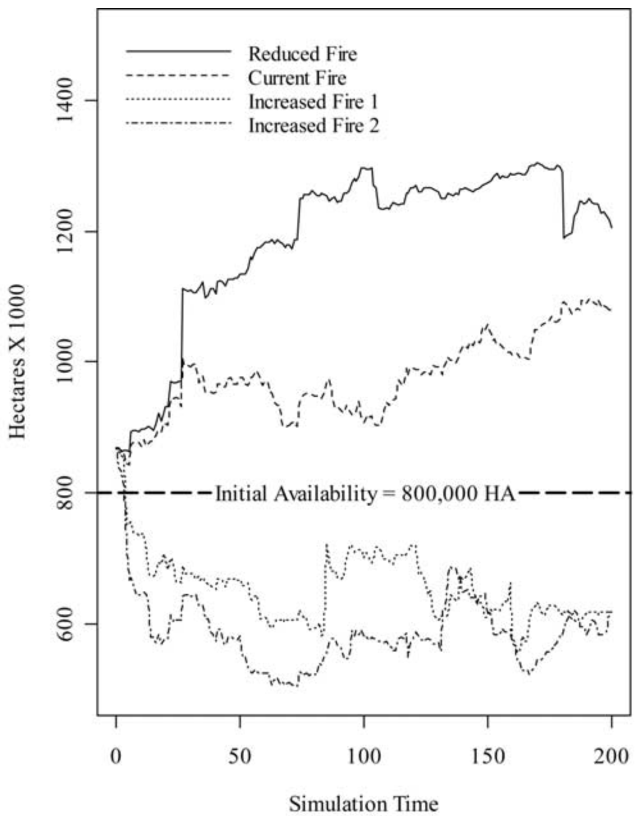

Figure 7. Time series showing the average amount of spruce forest on the landscape (where the spruce stand is >80 a old) for each fire frequency scenario. Results are from five replicates per scenario. The dashed line (800,000 ha) represents the amount of spruce-lichen forest initially available on the landscape used by the Nelchina Caribou herd near Fairbanks, Alaska [*Rupp et al.*, 2006]. Reprinted with permission from Ecological Society of America.

The number of wildfires in North America is increasing [*Kasischke and Stocks*, 2000]. Consistent data on how fire behavior might be changing, the return cycle, severity, or recovery patterns, are not available. Fires cause a short-term increase, but a long-term decrease, in near-surface soil moisture. After a fire, albedo may decline by 4%, resulting in increased shortwave input and surface temperature. The high surface temperature results in high long-wave radiation from the ecosystem to the atmosphere. Despite lower albedo following a fire, net radiation actually decreases [*Yoshikawa et al.*, 2002]. Meanwhile, transpiration decreases but evaporation increases. The immediate impacts of moderate fires include death of vegetation and increase in soil moisture and temperature. In cases of more severe fires, the surface organic layer may be entirely combusted, exposing the mineral soil beneath. Any disturbance to the surface layer will increase heat flow through the active layer into the permafrost. Approximately 3 to 5 a following a severe fire (depending upon site conditions), the active layer thickness increases, and talik formation and thermokarsting soon follow. Thicker talik and vegetation recovery (higher albedo and transpiration) will make the ground surface drier. This represents a ‘‘turning point’’ in the soil moisture regime.

[58] The dramatic effect of fire on soils significantly impacts the flora and fauna that inhabit an area. Following fire, some species recolonize over periods of 5–10 a; others take significantly longer, such as caribou, which may not return for 60 or more years following fire; indeed, preferred winter habitat for caribou may require up to a century to regenerate [*Rupp et al.*, 2006]. Caribou are most productive in spruce-lichen habitat more than 80 a old. Figure 7 shows simulations of the effect increased fire frequency might have on prime caribou habitat in the Nelchina Basin, Alaska. The two most likely increased fire scenarios show a net loss in caribou habitat that does not recover during a 200-a simulation period. Acute changes in species distribution obviously create a potential vulnerability for communities that rely on a particular species for a major part of their seasonal hunting or fishing. Historically, this may have been mitigated by the fact that many arctic communities were mobile and migrated to areas where desired species flourished. Establishment of more stationary communities has led to a greater reliance on resident, local subsistence resources that are vulnerable to climate-induced ecological change.

[59] Shrubs have been increasing in the tundra and appear to be favored over mosses and lichens in warmer climates [*Cornelissen et al.*, 2001; *Callaghan et al.*, 2004a], though there is also some evidence that changes in atmospheric ozone and carbon dioxide could make foliage more susceptible to damage by late spring or early autumn frost events [*Beerling et al.*, 2001]. Conversion of tundra to shrubs in warming climates invokes a positive feedback cycle [*Sturm et al.*, 2005]. However, wholesale movement of the boreal forest into present-day tundra is more complex, depending in part on seed dispersal mechanisms and how these might change [*Chapin et al.*, 2004], soil properties, and changes in hydrology [*Skre et al.*, 2002]. In addition, trees of a given species have shown differing growth trends in a warming climate [*Wilmking et al.*, 2004, 2005], and warmer summers could allow greater drying in drought years. Such behaviors indicate that arboreal response may not simply involve northward expansion.

[60] Even without increased shrubs, arctic snow cover may increase in some areas [*Hosaka et al.*, 2005]. Increased snow cover and ice crusting could impair the ability of animals to reach underlying vegetation [*Callaghan et al.*, 2004b], though it may also provide greater shelter to small animals that burrow underneath the snow [*Callaghan et al.*, 2004b]. Warming may also reduce lake ice thickness and cover, reducing winter survivability of fish [*Finstad et al.*, 2004]. Insects, on the other hand, may find more favorable conditions for wintering over in a warmer climate, and changes in circulation coupled with warming potentially could allow species to spread into new areas [*Coulson et al.*, 2002; *Callaghan et al.*, 2004b].

5.1.2. Rivers

[61] Many characteristics of rivers are expected to change with a changing climate, leading to impacts on the ecology of rivers that may in turn influence the well-being of humans who rely on the waterways. Increased river flow may affect temperature, sedimentation, and nutrient transport, which can affect habitat and breeding conditions, particularly for anadromous fish. Increased discharge may be associated with accelerated transport of materials that

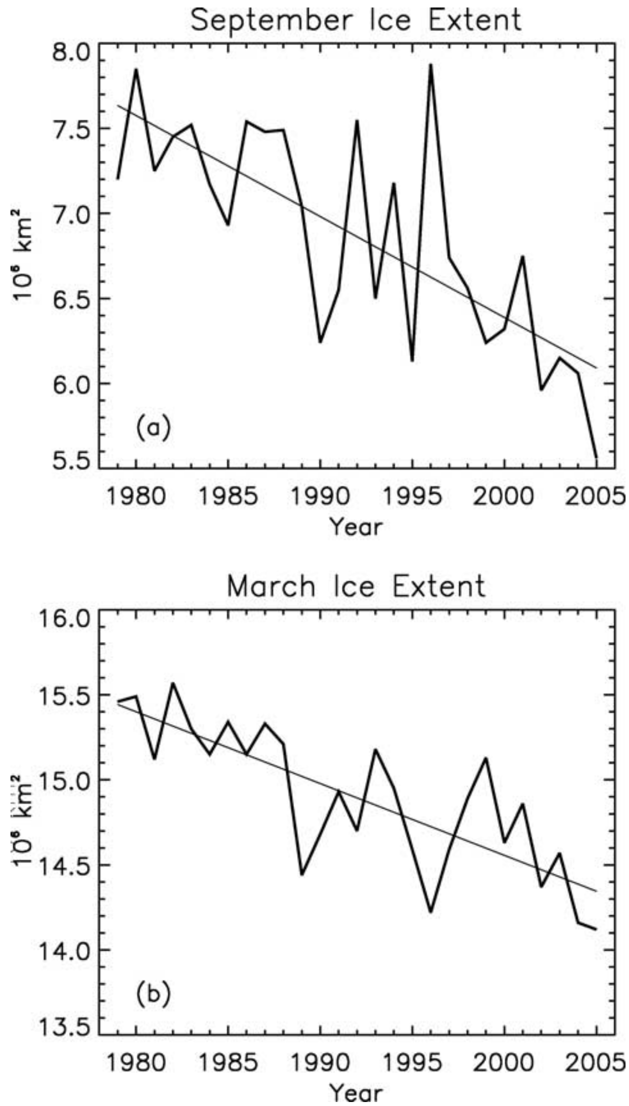

Figure 8. The time series of (a) September and (b) March ice extent from 1979 to 2005 [*Fetterer and Knowles*, 2002]. The thin line indicates the linear trend in the time series.

may affect water quality for biota, including those biota used for human consumption. This type of small-scale change can cause regional-scale vulnerabilities for many arctic residents who rely heavily on anadromous fish such as salmon and coregonid whitefish [*Huntington et al.*, 1999]. In addition, changes in the distributions of some species, such as beaver, will affect migration corridors for other species that periodically occupy the same niche (i.e., salmon blocked by beaver dams).

[62] Travel via rivers may be affected by changes in water level and its seasonality [*Fox*, 2002]. The effects of altered hydrology on fish populations are poorly understood, and river depth and water quality coupled with the seasonality of discharge are complicating factors. Timing of river freezing and thawing can affect both the recruitment dynamics of anadromous fish and the ability to harvest biota in desired stages of their life histories (e.g., fish with optimal fat content) [e.g., *Arctic Climate Impact Assessment* (*ACIA*), 2005].

5.2. Impacts of Arctic Atmospheric Change

[63] Many potential atmospheric changes lead to increases in average temperature [*Cassano et al.*, 2006], changes in the amount and type precipitation [*Cassano et al.*, 2006, 2007], and changes in tracks and intensity of storms [*Finnis et al.*, 2007]. Potentially significant changes in atmospheric composition may also occur, especially changes in methane [*Committee on the Science of Climate Change*, 2001], ozone [*Gillett et al.*, 2003], and carbon dioxide [*Johnson et al.*, 2002]. Each of these changes has been studied for impacts over the coming decades, but the ongoing and projected warming may be the single most important factor affecting ecosystems and human societies. Warming is also probably the most robustly projected change. Other climatic features may have important impacts, even if their trends are less discernable at present. Winter snow depth, for example, may increase in much of Siberia and northern Alaska [*Hosaka et al.*, 2005] because a warmer atmosphere would allow greater atmospheric moisture in regions that remain well below freezing despite winter warming [*Cassano et al.*, 2007]. Other impacts could conceivably involve atmospheric processes that occur more strongly and frequently in a warmer climate, such as intense, summertime convective storms. Thus most studies of climate impacts on arctic ecosystems and human societies focus on warming because it appears to be pervasive and highly likely, but other changes should not be ignored, even though they are less certain [*Simpson et al.*, 2002].

5.3. Impacts of Changes in Oceans and Sea Ice

[64] Ocean changes include physical changes, primarily reduction in sea ice extent and corresponding seasonal changes, and biological changes, including productivity and the distribution and abundance of various species (Figure 8). Although sea ice is unlikely to be a significant factor for terrestrial or near-shore freshwater budgets, associated biophysical changes will lead to societal impacts in various forms.

[65] Reduced sea ice will affect those who travel on, hunt from, or otherwise use sea ice as a platform. The timing, thickness, and patterns of breakup of sea ice have direct implications to a large range of biota. Access for boating will be improved, but spring marine mammal hunting may be hindered by reduced access or greater hazard owing to reduced ice thickness and extent [*ACIA*, 2005; *Gearheard et al.*, 2006]. Marine mammals dependent on sea ice, particularly polar bears and walrus, are currently experiencing direct impacts such as isolation (e.g., polar bears stranded on fast-melting ice or, conversely, on land), loss of access to prey species, and changes in the location of ice-associated phytoplankton and zooplankton and associated food webs. Sea ice provides access to human communities for subsistence activities, and unfamiliar patterns may render local knowledge unreliable under conditions of rapid change [*Gearheard et al.*, 2006]. At the same time, reduced sea ice may lead to greater ship traffic in the Arctic Ocean, both for transit from Asia to Europe and for access to arctic resources, including mineral resources as well as fish stocks [*ACIA*, 2005].

[66] Changes in sea ice extent and thickness will change light penetration into the upper ocean, change mixing of surface and deeper water, and change the location of zones of

freshwater input from melting ice. These changes will affect the distribution of species throughout the food web, with some species flourishing in the new conditions and others being reduced in abundance.

[67] Changes in productivity and species distribution will in turn affect people who use living marine resources. Combined with changes in sea ice, and the potential for increased ship traffic, greater access to marine resources may increase competition for those resources and also increase potential threats from pollution, disturbance, or accidents such as oil spills.

[68] While the precise trajectories of, and interactions among, these changes are hard to project, a likely scenario is an increase in human activity (and perhaps total population) in the region, increased exploitation of living and mineral resources (including oil and gas), and an overall shift to conditions more typical of subarctic seas today.

6. Summary

[69] Changes to the terrestrial freshwater cycle in the Arctic are intimately tied to permafrost. As permafrost warms, the active layer thickens, and thermokarst formation occurs. River discharge may change in ways not predictable from precipitation and evapotranspiration alone owing to changes in storage. Lakes may be expected to appear and grow in areas of continuous permafrost, while shrinking and disappearing in more degraded permafrost. A deeper active layer will have the capacity to attenuate peak flows of runoff owing to increased storage, and thermokarst geomorphology will create more heterogeneity in soil moisture, leading to ponding in some places but drying in other areas. Complex feedbacks are expected, such as conditions more conducive to fire, which then lowers albedo and initially increases soil moisture.

[70] It appears that the storage of water vapor in the arctic atmosphere has increased along with the poleward transport of moisture. Additional open water resulting from reduced sea ice extent also provides a moisture source for the arctic atmosphere. All these factors should be expected to contribute to increased arctic net precipitation as well, but available data cannot yet indicate definitively whether this has occurred. It is clear, however, that varying large-scale circulation patterns have a strong influence on all three components of the atmospheric hydrologic system. A predominantly positive phase of the NAO/AO index, in effect since the late 1980s, leads to an apparent acceleration of the arctic water cycle. Further investigation with recently improved corrections to gauges [*Cherry et al.*, 2005]; longer and better satellite retrievals and more realistic reanalysis products [*Serreze et al.*, 2006] should lead to a better understanding of this important component of the global climate system.

[71] Changes in ocean circulation systems and thermodynamics have been observed and are expected to continue. Sea ice has decreased in summer extent and thinned, and projections indicate that losses will continue. As the oceans receive different amounts of freshwater discharge from rivers, the circulation, chemistry, and ecology will likely change, which in turn will affect conditions on land, coastal water quality, and sea life, all of which have major impacts on fisheries and subsistence food supply.

[72] Components of the arctic freshwater system affect and are used by people and ecosystems in various ways. Changes in the arctic freshwater system and its cycle are likely to have a range of effects; some of these may be sensitive to the magnitude and direction of specific changes and subject to complex interactions within and between the biophysical and societal realms. Communities and individuals will need to anticipate changes and their impacts in order to respond effectively. Further study is needed to examine the details of human-freshwater interactions, the likely consequences of change, and the ways in which one must adapt in a changing climate.

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