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Key Points:
- Specific humidity and precipitation have generally increased in the Arctic
- The Arctic water cycle is projected to intensify during this century
- Complexity of physical processes causes challenges in modeling of water cycle

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The atmospheric role in the Arctic water cycle: A review on processes, past and future changes, and their impacts

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Abstract Atmospheric humidity, clouds, precipitation, and evapotranspiration are essential components of the Arctic climate system. During recent decades, specific humidity and precipitation have generally increased in the Arctic, but changes in evapotranspiration are poorly known. Trends in clouds vary depending on the region and season. Climate model experiments suggest that increases in precipitation are related to global warming. In turn, feedbacks associated with the increase in atmospheric moisture and decrease in sea ice and snow cover have contributed to the Arctic amplification of global warming. Climate models have captured the overall wetting trend but have limited success in reproducing regional details. For the rest of the 21st century, climate models project strong warming and increasing precipitation, but different models yield different results for changes in cloud cover. The model differences are largest in months of minimum sea ice cover. Evapotranspiration is projected to increase in winter but in summer to decrease over the oceans and increase over land. Increasing net precipitation increases river discharge to the Arctic Ocean. Over sea ice in summer, projected increase in rain and decrease in snowfall decrease the surface albedo and, hence, further amplify snow/ice surface melt. With reducing sea ice, wind forcing on the Arctic Ocean increases with impacts on ocean currents and freshwater transport out of the Arctic. Improvements in observations, process understanding, and modeling capabilities are needed to better quantify the atmospheric role in the Arctic water cycle and its changes.

1. Introduction

The atmosphere contains water in the forms of vapor, liquid, and ice. The total water content in the atmosphere is approximately 13,000 km^3 [*Gleick*, 1996], of which 200 km^3 over the Arctic [*Serreze et al.*, 2006]. These numbers are very small compared to the ocean, ice sheets, glaciers, lakes, rivers, and ground. However, the atmosphere is a fundamental component of the water and energy cycles. Water vapor in the Arctic atmosphere has a residence time of about a week, compared to a decade for freshwater in the Arctic Ocean [*Carmack et al.*, 2016], and thousands of years for ice sheets and glaciers. Atmospheric moisture, clouds, and precipitation simultaneously affect and are affected by the recent rapid climate change in the Arctic.

Our understanding of the past and present-day Arctic and high-latitude hydrological cycle is based on in situ and remote sensing observations, atmospheric reanalyses, and climate model hindcasts. Longer-term historical context is provided by proxy (paleo) indicators. Estimates on future hydrological changes are derived from climate model projections with various scenarios for human activities and associated external forcings. However, understanding of the atmospheric moisture budget in the Arctic is far from complete. The main challenges are primarily related to sparcity (in space and time) of accurate in situ and remote sensing observations. This is the case for almost all moisture variables, including atmospheric specific and relative humidity, clouds (coverage, type, layers, phase, and properties), precipitation (amount and phase), and evapotranspiration. In many cases, the uncertainties are largest over sea ice and the open ocean but considerable also over land areas. Also atmospheric reanalyses and climate model results include

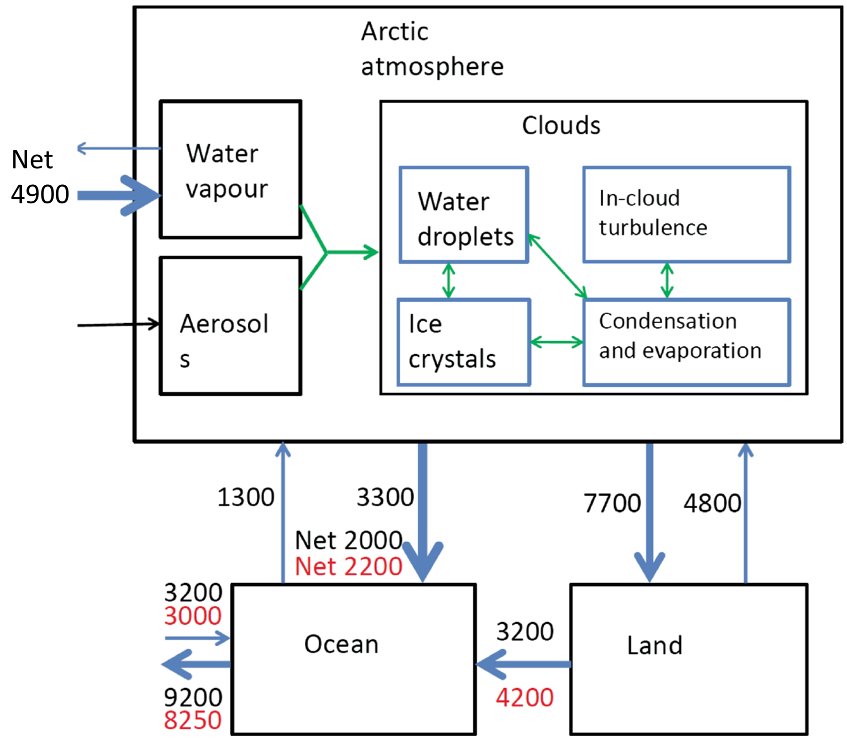

Figure 1. Schematic presentation of the main freshwater components and processes in the Arctic atmosphere. The numbers in black indicate the freshwater transports in km^3/yr on the basis of *Serreze et al.* [2006], representing the period of 1979–2001, whereas the numbers in red are estimated for the period 2000–2010 by *Haine et al.* [2015].

large errors and uncertainties in the Arctic [*Jakobson et al.*, 2012; *Liu et al.*, 2015; *Lique et al.*, 2016]. In the case of reanalyses, these are affected by the lack of good observations and limited consideration on homogeneity of the data throughout the entire analysis period using the state-of-the-art assimilation methods [*Trenberth*, 2016; *Zhang et al.*, 2013]. An important source of errors, for both reanalyses and climate models, is the suite of complex physical processes extant in the Arctic atmosphere and their interaction with the Earth surface [*Vihma et al.*, 2014]. These processes include, for example, physics of mixed-phase clouds, radiation-cloud-turbulence interactions, and evaporation from drifting/blowing snow and spray droplets over the open ocean.

Despite the scarcity of data, efforts were made as early as the 1960s to evaluate the freshwater budget of the Arctic Ocean [*Mosby*, 1962; *Aagaard and Greisman*, 1975; *Carmack et al.*, 2016]. *Budyko* [1963] presented estimates on global evaporation, and *Palmen and Vuorela* [1963] first calculated atmospheric water vapor transport, in the region −40°S to 40°N, soon followed by estimates reaching as far north as 70–75°N [*Starr et al.*, 1965; *Rakipova*, 1966; *Oort*, 1971]. The first Pan-Arctic moisture budget estimates were made by *Oort* [1975]. A recent synthesis of the large-scale freshwater cycle of the Arctic, including its atmospheric, oceanic, and terrestrial components during 1979–2001, was presented by *Serreze et al.* [2006]. Although the study was more focused on the ocean than atmosphere, the results for the latter included important estimates for the mean values, annual cycles, and interannual variations for the total (i.e., vertically integrated) water vapor (TWV), evapotranspiration (i.e., the sum of evaporation from the Earth surface and transpiration from plants), precipitation, and net precipitation (i.e., precipitation minus evapotranspiration) over the Arctic sea and land areas (Figure 1). A more recent major effort was the assessment "Snow, Water, Ice, and Permafrost in the Arctic" (SWIPA), which included a review on recent variations in the Arctic climate [*Walsh et al.*, 2011a]. The important message was that the years since 2000 have been quite wet according to both precipitation and river discharge data. There are also indications of increases in cloudiness over the Arctic, especially in low clouds during the warm season [*Walsh et al.*, 2011a, 2011b]. In the SWIPA assessment, however, atmospheric moisture, clouds, and precipitation were only addressed among the many other climate variables without any particular focus on the atmospheric freshwater system in the Arctic. A recent synthesis on the intensification of the Arctic freshwater cycle was presented by *Rawlins et al*. [2010] and its feedbacks and impacts on terrestrial, marine, and human life were reviewed by *Francis et al*. [2009]. In Figure 1, we schematically illustrate the Arctic freshwater budget with a focus on the atmosphere; with increases in river runoff and net precipitation, also this figure demonstrates a recent intensification of the freshwater cycle.

The objectives of this paper are (a) to briefly review the role of Arctic atmospheric moisture, clouds, evaporation, and precipitation in the Earth system, (b) to provide an up-to-date quantification of the historical and ongoing changes in atmospheric freshwater in the Arctic and their drivers, (c) to evaluate the nature and causes of changes expected during the 21st century, (d) to provide a synthesis of the impacts of the changes, and (e) to identify the knowledge gaps. This paper is a contribution to the "Arctic Freshwater Synthesis (AFS)" (http://www.climate-cryosphere.org/activities/targeted/afs and *Prowse et al*. [2015a] which is a joint effort to review the current knowledge on the Arctic freshwater system and its components: the ocean [*Carmack et al*., 2016], terrestrial hydrology [*Bring et al*., 2016], ecosystems [*Wrona et al*., 2016], and water resources [*Instanes et al*., 2016], with a separate synthesis on modeling of the system [*Lique et al*., 2016]. Following the AFS introductory paper [*Prowse et al*., 2015a], we define the Arctic domain as the area consisting of the Arctic Ocean, its marginal seas, and the land areas contributing to river discharge into these seas. Precipitation and evaporation in the river basins are important components of the atmospheric role in the Arctic water cycle. However, not all literature we cite has used the same definition of the Arctic. Hence, we point out the different definitions, when the differences affect the conclusions made.

2. System Function and Key Processes

The spatial and temporal distributions of water vapor in the Arctic are closely related to air temperature distributions. The moisture holding capacity of air increases exponentially with air temperature according to the Clausius-Clapeyron equation. Hence, TWV (or precipitable water) decreases poleward, decreases from summer to winter, and is higher over the open ocean than sea ice and cold continents (Figure 2) (see *Webster* [1994] for thorough discussion on the climatological importance of the Clausius-Clapeyron equation). In the Arctic, clouds are very common over the open ocean and sea ice, but less common over the continents. The cloud fraction reaches its maximum in late summer and early autumn and a minimum during late winter [*Curry et al*., 1996; *Shupe et al*., 2011]. In the central Arctic, typical winter conditions swap between overcast and clear skies [*Makshtas et al*., 1999] determined mostly by the large-scale atmospheric circulation [*Morrison et al*., 2012]. In summer low stratus or stratocumulus clouds are present for most of the time [*Curry et al*., 1996; *Tjernström et al*., 2012]. The occurrence of water vapor and clouds in the Arctic is partly due to local evaporation/evapotranspiration and condensation and partly due to transport from lower latitudes (Figure 1). The transport is mostly due to transient cyclones [*Jakobson and Vihma*, 2010]. Information on precipitation over the central Arctic is limited and of questionable accuracy, but atmospheric reanalyses suggest that the large-scale spatial and seasonal distributions of precipitation roughly follow those of TWV (Figure 2), but with added spatial detail due to orography. As an annual mean, precipitation exceeds evapotranspiration over almost all land and ocean areas in the Arctic and sub-Arctic, the primary exceptions being parts of the Norwegian and Barents Seas [*Jakobson and Vihma*, 2010]. In the following subsections we provide more detailed information on the different aspects of atmospheric water cycle in the Arctic.

2.1. Atmospheric Moisture

Water vapor is the source for cloud and fog formation, but also affects radiative transfer and evaporation/condensation. Relative humidity, together with the atmospheric dynamics and the availability and properties of condensation nuclei, controls the cloud formation. The level where saturation is reached depends on the vertical profiles of temperature and specific humidity. Inversions, i.e., layers were the values increase with height, are common in the Arctic for air temperature as well as specific and relative humidity. The air temperature inversions occur somewhere between the surface and 1–1.5 km altitude throughout the year but display a strong seasonality [e.g., *Serreze et al*., 1992; *Tjernström and Graversen*, 2009]. Also, specific humidity inversions have a high importance in the Arctic climate system, especially for cloud formation

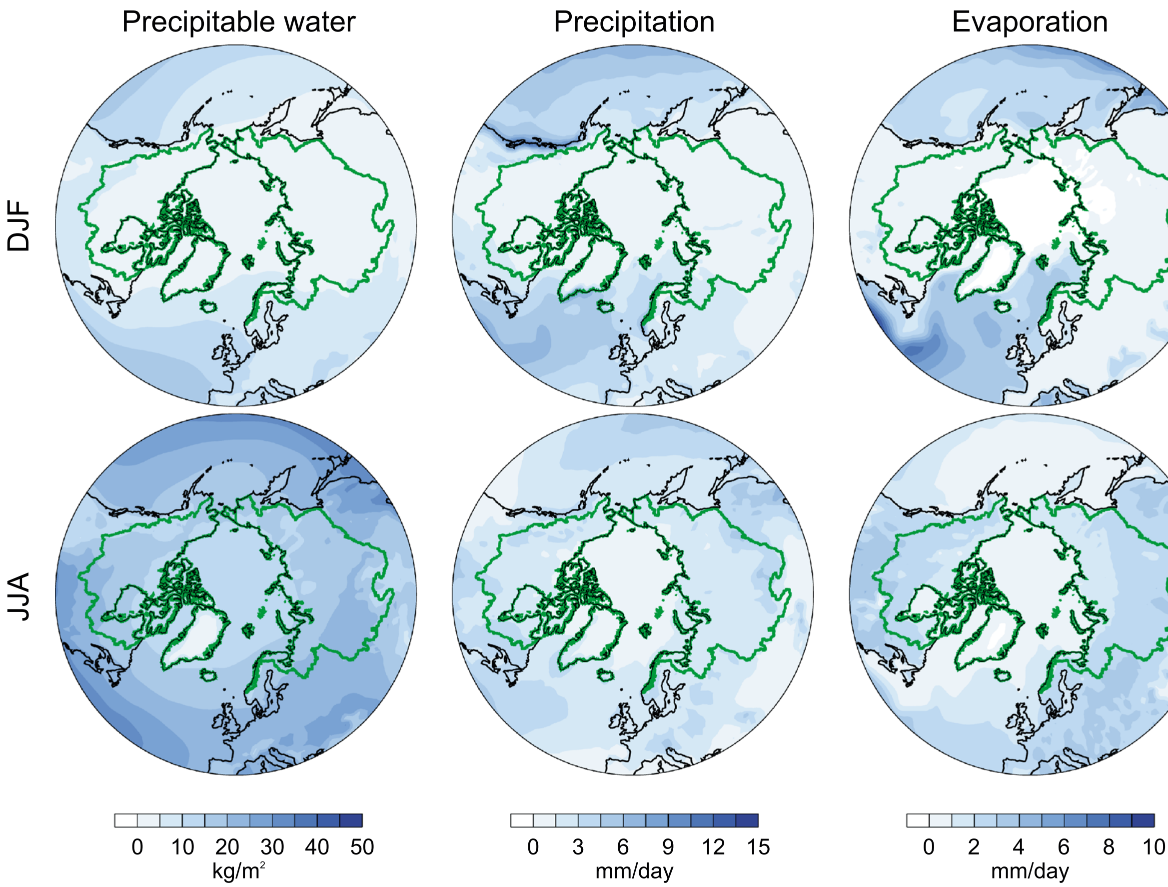

Figure 2. Seasonal means of precipitable water, precipitation, and evaporation for winter (DJF) and summer (JJA) during the period 1980–2013 on the basis of JRA-55 reanalysis. The green lines indicate the boundaries of the Arctic river catchment.

and maintenance, across a wide range of spatial and temporal scales [*Nygård et al*., 2014]. Specific humidity inversions are often collocated with temperature inversions, with the base near the temperature inversion base close to the cloud top. Hence, during entrainment (mixing of the cloud layer air with the warmer inversion layer air above it) the inversion layer acts as a moisture source to the cloud layer [*Solomon et al*., 2011; 2014; *Sedlar et al*., 2012]. In the case when cloud layers are decoupled from the Earth surface (due to a secondary inversion below the cloud base), moisture inversions may be the only moisture source to the clouds [*Solomon et al*., 2011; 2014; *Savré et al*., 2015]. Even a small vertical gradient in specific humidity may be important for the occurrence of condensation or for the rate of evaporation from existing cloud droplets or ice crystals [*Nygård et al*., 2014].

In addition to effects on clouds, the profile of specific humidity is important for the longwave and shortwave radiative transfer in the atmosphere. The humidity profile matters particularly for longwave radiation in clear-sky conditions [*Devasthale et al*., 2011], when the difference in downward longwave radiation at the Earth surface between dry and humid conditions may reach 10–20 W/m^2 in conditions typical for the Arctic [e.g., *Prata*, 1996]. The effect is enhanced by the common co-occurrence of temperature and humidity inversions in the Arctic; humid air emits more longwave radiation than dry air because of its humidity and high temperature.

The loss of sea ice and snow cover, and increases in atmospheric moisture and clouds, amplify warming in the Arctic region [e.g., *Manabe and Stouffer*, 1980; *Screen and Simmonds*, 2010a, 2010b; *Pithan and Mauritsen*, 2014]. As a consequence, the Arctic warms faster than the middle latitudes (or global average), a phenomenon known as Arctic amplification [*Serreze and Francis*, 2006; *Serreze et al*., 2009; *Serreze and Barry*, 2011]. As water vapor is the strongest greenhouse gas, an increase in its content yields warming at all latitudes. By itself it does not cause the Arctic amplification, but it has been suggested to amplify the feedbacks due to clouds, surface albedo, and air temperature (the lapse rate and Planck feedbacks) [*Pithan and Mauritsen*, 2014]. Warming generated by these feedbacks yields increased water vapor contents and associated radiative effects, which further enhance the Arctic amplification [*Langen et al.*, 2012].

2.2. Evaporation and Moisture Transport From Lower Latitudes

Atmospheric moisture in the Arctic is a result of local evaporation and moisture transport from lower latitudes (Figure 1). Due to large evaporation from leads and polynyas, the near-surface air humidity over Arctic sea ice is usually at or close to saturation with respect to ice phase [*Andreas et al.*, 2002], and the sublimation from snow/ice surface is therefore weak [e.g., *Persson et al.*, 2002]. Sublimation from drifting/blowing snow may be more efficient because of the large surface area of snowflakes in the air [*Déry and Tremblay*, 2004]. Part of the drifting/blowing snow is transported to leads, representing a freshwater flux to the ocean [*Fichefet and Morales Maqueda*, 1999]. Compared to sea ice, sublimation from drifting/blowing snow is probably more important over Arctic and sub-Arctic land areas, where the air is drier.

Over the open ocean under strong winds evaporation from spray droplets gives an important addition to evaporation from the ocean surface. According to *Andreas et al.* [2008], with a wind speed of 11–13 m s^{-1} evaporation from spray droplets is 10% or greater of the surface evaporation, and with a wind speed of 25 m s^{-1}, common for Polar lows (intensive mesoscale cyclones in Polar regions) [e.g., *Rasmussen and Turner*, 2003], the contribution is already of the order of 70–80% (for sea surface temperature (SST) = 0°C). Under such wind speeds, however, the numbers include a considerable uncertainty due to lack of data. The process of evapotranspiration from vegetated surfaces is thoroughly discussed in *Bring et al.* [2016], and therefore not addressed here.

Another source for atmospheric moisture in the Arctic is transport from lower latitudes, driven by the north-south gradient in air specific humidity and affected by large-scale circulation patterns, such as planetary waves, the subtropical jet stream, the Polar front jet stream, storm tracks, as well as the Halley, Ferrel, and Polar cells. In a classical approach [*Palmen and Vuorela*, 1963] the transport is dived into contributions of mean meridional circulation, stationary eddies, and transient eddies. Stationary eddies represent deviations from the zonal mean, whereas transient eddies represent deviations from the temporal mean, i.e., transient cyclones. On the basis of ERA-40 reanalysis, *Jakobson and Vihma* [2010] calculated that transient eddies dominate the moisture transport across 70°N, contributing to 80–90% of the total northward transport. The contribution of mean meridional circulation is only 1% in summer to 12% in winter. Changes in poleward moisture transport may originate from two reasons: changes in north-south moisture gradient and atmospheric circulation. The first is expected to dominate in time scales of several decades and longer [*Skific et al.*, 2009], whereas the second dominates in interannual and shorter time scales. According to *Sorteberg and Walsh* [2008], the interannual variability in moisture transport is mainly driven by variability in cyclone activity over the Greenland Sea and East Siberian Sea. Regional linkages are complex. For example, if many cyclones enter the central Arctic from the Kara Sea, it reduces the total moisture transport. This is because after reaching the central Arctic, the Kara Sea cyclones generate northerly wind anomalies over the Greenland and Barents Seas, reducing the moisture transport from there [*Sorteberg and Walsh*, 2008]. Atmospheric rivers (narrow corridors of intensive transport, which occur within the warm conveyor belt of extratropical cyclones) are a key component of the poleward moisture transport in winter, in particular in events of extreme transport [*Newman et al.*, 2012; *Gimeno et al.*, 2014; *Neff et al.*, 2014].

The sensitivity of the above mentioned results to the reanalysis applied is not known. Further, the vertical distribution of moisture transport is poorly known. On the basis of rawinsonde data, *Overland and Turet* [1994] found out that the poleward transport across 70°N peaks at the 850 hPa level, but *Jakobson and Vihma* [2010] detected the peak at 930 hPa level in winter and at 970–990 hPa level in other seasons. Reanalyses are liable to errors in moisture variables [e.g., *Jakobson et al.*, 2012], but sea areas are underpresented in circumpolar estimates based on radiosonde data.

2.3. Clouds

Clouds have a large impact on climate through interaction with radiation but, due to lack of in situ observations and complexities in monitoring clouds from space (section 6), the cloud climatology for the Arctic is far from complete. Clouds also remain one of the largest uncertainties for projections of climate change [*Lique et al.*, 2016, section 4]. Several Arctic characteristics contribute to clouds having Arctic-specific effects. These include high-reflective ice/snow-covered surfaces and the pronounced seasonal cycle in solar radiation, where the Sun is absent for a large part of winter and present all day in summer, but at large solar zenith angle.

Available data indicate that the Arctic is very cloudy, with a pronounced annual cycle in cloudiness, ranging from 40–70% in winter to 80–95% in summer and autumn [*Intrieri et al.*, 2002b; *Shupe et al.*, 2011]. Low clouds and fog dominate [*Intrieri et al.*, 2002a, 2002b; *Shupe et al.*, 2011; *Shupe*, 2011], predominantly having a warming effect on the surface, especially over the ice-covered Arctic Ocean. In winter, with no solar radiation, clouds determine the distribution of net thermal radiation (defined positive downward), a distinct bimodal distribution with one peak at ~ -40–50 W m^{-2}, with the lowest temperatures and associated with clear conditions, and another one near 0 W m^{-2} [*Stramler et al.*, 2011; *Morrison et al.*, 2011] associated with clouds. Also through a large part of summer, longwave radiation dominates the cloud radiative effect (CRE), although solar radiation is present; this is due to a high surface albedo and a large solar zenith angle [*Sedlar et al.*, 2011]. Except for, possibly, July and August, when open water and melt ponds are present, clear-sky conditions reduce surface net longwave radiation more than they increase the net shortwave radiation [*Curry et al.*, 1996; *Intrieri et al.*, 2002a; *Shupe and Intrieri*, 2004].

Cloudy conditions are often related to a warmer free troposphere, as cloudy episodes are typically associated with advection from lower latitudes [*Vihma and Pirazzini*, 2005; *Morrison et al.*, 2011]. Cloud base temperatures are sometimes higher than the snow/ice surface temperature, enhancing CRE [*Persson et al.*, 2002; *Persson*, 2012]. However, if the bases of temperature and humidity inversions are located close to the cloud top, the higher temperature and specific humidity aloft do not affect CRE; their radiative effects at the surface are masked by the presence of optically thick clouds [*Tjernström et al.*, 2012]. Near the marginal ice zone, strong sustained surface inversions may form in cases of on-ice flow; in these cases specific humidity may also increase with altitude, or low clouds or fog may form. The inflow of this warm and moist air has a substantial impact on the radiative flux both at the surface and at the top of the atmosphere [e.g., *Tjernström et al.*, 2015].

Clouds also control the vertical structure of the lower troposphere. Contrary to shear-driven near-surface turbulence [*Persson et al.*, 2002; *Tjernström et al.*, 2012] and convection over leads, polynyas and the open ocean in winter [*Lüpkes et al.*, 2012], much of the mixing in the lower troposphere over sea ice is driven by cloud overturning from cloud top radiative cooling. This mixing sometimes contributes to forming a single deep well-mixed layer [*Vihma et al.*, 2005], but more often the clouds are decoupled from the surface [*Tjernström*, 2007; *Shupe et al.*, 2013; *Sotiropoulou et al.*, 2014]. A key to understanding these effects lies in the cloud microphysics and interactions with aerosols. Measurements from expeditions into the Arctic indicate that aerosol concentrations in the Arctic are low compared to the rest of the Earth, especially in summer [*Tjernström et al.*, 2012, 2014]. Paucity of ice-forming particles favors mixed-phase clouds, common in the Arctic even at very low temperatures [*Prenni et al.*, 2007]. These have a thin supercooled liquid water layer at the cloud top, continuously shedding frozen precipitation (see *Morrison et al.* [2011] for a review). The net effect is persistent clouds [e.g., *Shupe*, 2011] with more or less constant but weak frozen precipitation and with a large impact on the surface energy balance [*Persson et al.*, 2002; *Tjernström*, 2005; *Prenni et al.*, 2007; *Sedlar et al.*, 2011; *Bennartz et al.*, 2013].

Low concentrations of cloud condensation nuclei (CCN) generate clouds with few but large droplets, and hence a relatively low cloud albedo [*Twomey*, 1977]. With exceptionally low concentration, droplets grow large enough to deposit. Since each droplet brings with it the CCN it formed on, this feeds back to an even lower CCN concentration and generates optically thin clouds [*Mauritsen et al.*, 2011]. Observations from several expeditions indicate that in summer this might occur as often as 30% of time. When CCN $<$ ~10 cm^{-3}, increasing CCN concentrations lead to a large increase in the longwave surface CRE, but a relatively small effect on solar surface CRE; the latter (the so-called Twomey effect) becomes important only for CCN $>$ 10 cm^{-3}, as the cloud becomes optically thick and emits as a blackbody. Increasing the aerosol concentrations may thus have both cooling and warming effect on the surface.

Studies indicate that the strong influence by clouds on CRE has a strong influence on the sea ice melt. *Kapsch et al.* [2013, 2014] indicate that years with more clouds and water vapor in spring from advection from south are the years with the lowest sea ice extent in September. Moreover, modeling studies [*Kapsch et al.*, 2015] also show a significant feedback; the lower sea ice concentration in autumn increases the cloudiness and causes thinner ice at lower concentration at the start of winter.

2.4. Precipitation

Quantitative estimates for annual accumulated precipitation and evaporation over the Arctic Ocean and terrestrial Arctic are shown in Figure 1 and their spatial distributions in Figure 2. In winter the spatial distributions of precipitation and evaporation are dominated by the difference between large values over the open seas and low values over snow/ice-covered sea and land areas (Figure 2). In summer, however, both precipitation and evaporation over Arctic land areas, except Greenland, exceed the values over sea areas north of 70°N and are comparable to those at lower latitudes. Precipitation and evaporation/evapotranspiration affect the ocean and terrestrial freshwater budgets, the surface albedo and energy budget, and the mass balance of ice sheets, glaciers, and sea ice. Over the Arctic Ocean, net precipitation is strongly positive (by 2000–2200 $km^3 yr^{-1}$) and, hence, freshens the ocean (Figure 1). Precipitation exceeds evaporation particularly during winter (Figure 2), when the surplus is stored on snow on top of sea ice. Hence, part of the Arctic Ocean freshening occurs during the snow and ice melt in summer. An even more important freshening factor is river discharge, which results from a positive net precipitation over the surrounding continents, especially Eurasia [*Serreze et al*., 2006; *Bring et al*., 2016]. Both net precipitation and river discharge have strong seasonal cycles. Over continents, in most of the Arctic river catchment, net precipitation is stored as snow for half a year or more. Hence, river discharge peaks in late spring during and soon after the snowmelt period [*Bring et al*., 2016], with 60% of the river discharge to the Arctic Ocean occurring from April through July [*Lammers et al*., 2001]. Net precipitation over the Arctic Ocean reaches its maximum later, typically in late summer and early autumn [*Walsh et al*., 1994]. This is partly due to the summer maximum in Arctic cyclone activity [*Zhang et al*., 2004; *Serreze and Barrett*, 2008].

The snow and ice surface albedo depends on several factors, such as the snow wetness, temperature, density, grain structure, impurities, melt ponds, and surface slope [e.g., *Pirazzini*, 2004; *Perovich et al*., 2009; *Perovich and Polashenski*, 2012]. The effects of precipitation on snow and ice albedo have received relatively little attention, although *Persson* [2012] estimated the rain effect of lowering surface albedo by approximately 0.07 at melt onset, and *Sedlar et al*. [2011] and *Persson* [2012] partly attributed the onset of the ice surface freeze in autumn to an albedo change due to new snow at a time when solar radiation is already decreasing. If precipitation falls as rain instead of snow, it has a dramatic effect on surface albedo and sea ice melt (see section 3.1.3 on detected changes in the phase of precipitation).

Precipitation is the only major source term for the mass balance of the Greenland ice sheet [*Bring et al*., 2016], ice caps, and glaciers in the Arctic. Most precipitation is generated orographically when moist air masses, typically related to cyclones, rise over the coastal slopes [*Schuenemann et al*., 2009]. Hence, roughly 2–3 times more precipitation falls over the coastal regions of Greenland than over the interior of the ice sheet, where low temperatures do not allow large amounts of precipitable water [*Hakuba et al*., 2012]. Precipitation also affects the mass balance of sea ice, but different effects partly compensate for each other. On the one hand, snowfall on top of sea ice enhances the thermal insulation and increases the surface albedo, the first effect reducing sea ice growth in winter [*Leppäranta*, 1993] and the latter reducing melt in spring and summer [*Cheng et al*., 2008]. On the other hand, snowfall on sea ice enhances sea ice growth via snow-to-ice transformation, which may take place via snow-ice formation due to flooding of sea water into the snow layer and subsequent refreezing of slush [e.g., *Kawamura et al*., 2001] and superimposed ice formation due to refreezing of meltwater or slush on top of sea ice [*Granskog et al*., 2006]. Although rain strongly favors sea ice melt, it may also provide a source for superimposed ice formation, temporally favoring sea ice growth. Although flooding is not yet common in the Arctic, future perspectives with thinner sea ice and increasing precipitation (section 4.1) suggest a potentially increasing percentage of snow ice and superimposed ice in the Arctic sea ice thickness. The above mentioned processes are relevant also for lake and river ice [*Bring et al*., 2016].

2.5. Summary

There is a multitude of complex processes related to air moisture, clouds, precipitation, and evaporation in the Arctic. The atmospheric processes closely interact with hydrological and oceanographic processes. It is

Table 1. Estimates of Trends for Atmospheric Freshwater Variables Over the Arctic Ocean and Its Drainage Basin[a]

Variable	Season	Trend	Regions and References
Near-surface specific humidity	Annual	I	Pan-Arctic (*Dai* [2006], *Vincent et al.* [2007], *Willett et al.* [2008, 2013], *Serreze et al.* [2012], and *Hartmann et al.* [2013])
	Summer and early autumn	I	Arctic Ocean (*Screen and Simmonds* [2010a] and *Serreze et al.* [2012]), the Norwegian, Barents, and Greenland Seas (*Serreze et al.* [2012]), eastern Canada (*Vincent et al.* [2007], *Serreze et al.* [2012], and *Gill et al.* [2013]), and eastern Eurasia (*Willett et al.* [2008] and *Gill et al.* [2013])
	Other seasons	I	Pan-Arctic except for localized decreases in Canada, Siberia, and the North Greenland Sea (*Vincent et al.* [2007] and *Willett et al.* [2008])
Cloud coverage	Annual	I and D	I: Pan-Arctic terrestrial regions north of 60°N, (*Eastman and Warren* [2013]), Siberia (*Nahtigalova* [2013], see also *Eastman and Warren* [2010], and *Walsh et al.* [2011a, 2011b])
			D: Eurasia north of 70–75°N (*Baikova et al.* [2002] and *Nahtigalova* [2013])
	Winter	I and D	I: Arctic Ocean (*Eastman and Warren* [2010])
			D: Arctic north of 60°N, particularly Arctic Ocean (*Liu et al.* [2008])
	Spring	I and D	I: Pan-Arctic north of 60°N, particularly eastern Canada and the Arctic Ocean (*Liu et al.* [2008])
			D: Pan-Arctic north of 70°N (*Screen and Simmonds* [2010a]) and Arctic Ocean (*Eastman and Warren* [2010])
Precipitation	Annual	I	Pan-Arctic land areas (*McBean et al.* [2005], *Min et al.* [2008], *Rawlins et al.* [2010], *Walsh et al.* [2011a, 2011b], *Hartmann et al.* [2013]), northern Canada (*Zhang et al.* [2000, 2011], *Mekis and Vincent* [2011], and *Linton et al.* [2014]), western Eurasia (*Hartmann et al.* [2013]), and a large part of Russia (*Roshydromet Federal Service for Hydrometeorology and Environmental Monitoring* [2014])
	Winter	I and D	I: Central Arctic Ocean (*Radionov et al.* [2013]), western Siberia (*Roshydromet Federal Service for Hydrometeorology and Environmental Monitoring* [2008]), southeastern Canada (*Zhang et al.* [2000] and *Mekis and Vincent* [2011]), coastal regions of Canadian Arctic (*Zhang et al.* [2000], *Mekis and Vincent* [2011], and *Linton et al.* [2014]), and Scandinavia (*Liston and Hiemstra* [2011])
			D: Northeastern Russia (*Roshydromet Federal Service for Hydrometeorology and Environmental Monitoring* [2008]) and midlatitude headwater parts of the basin of Mackenzie River (*Zhang et al.* [2000], *Mekis and Vincent* [2011], *Yip et al.* [2012], and *Linton et al.* [2014])
	Spring and summer	I	Canada, but in summer both I and D in the upper McKenzie River basin (*Zhang et al.* [2000], *Mekis and Vincent* [2011], *Zhang et al.* [2011], *Yip et al.* [2012], *Linton et al.* [2014])
	Summer and autumn	I and D	I: East Siberia (*Roshydromet Federal Service for Hydrometeorology and Environmental Monitoring* [2008])
			D: Central Eurasia (*Rawlins et al.* [2006] and *Bogdanova et al.* [2010])
Snowfall fraction	Winter	I	Northern Canada (*Zhang et al.* [2000], *Vincent and Mekis* [2006], *Liston and Hiemstra* [2011], *Zhang et al.* [2011]) and midlatitude Eurasia (*Liston and Hiemstra* [2011]), particularly over the north of Yenisey and Ob River basins (*Bogdanova et al.* [2010])
	Primarily spring and autumn	D	Southern Canada (*Zhang et al.* [2000, 2011], *Vincent and Mekis* [2006], *Linton et al.* [2014]) and high-latitude Eurasia south of Siberia (*Bogdanova et al.* [2010] and *Liston and Hiemstra* [2011])
	Summer	D	Arctic Ocean (*Screen and Simmonds* [2012])

Table 1. (continued)

Variable	Season	Trend	Regions and References
No. of days with heavy precipitation	Annual	I and D	I: western Eurasia (*Alexander et al.* [2006] and *Donat et al.* [2013]) and northern Canada (*Zhang et al.* [2001] and *Vincent and Mekis* [2006]). For medium/heavy snowfall: Eurasia (*Borzenkova and Shmakin* [2012])
			D: western Canada (*Alexander et al.* [2006])
Daily *P* intensity	Annual	I and D	I: northern Canada (*Vincent and Mekis* [2006], *Peterson et al.* [2008], and *Donat et al.* [2013]) and Eurasia (*Donat et al.* [2013])
			D: southern Canada (*Vincent and Mekis* [2006], *Peterson et al.* [2008], and *Donat et al.* [2013]) and coastal northern Russia (*Donat et al.* [2013]),
No. of consecutive dry days	Annual	D	western Canada (*Alexander et al.* [2006] and *Vincent and Mekis* [2006]) and western Eurasia (*Donat et al.* [2013]).
Evapotranspiration	Various	I	Annually most of Canada (*Burn and Hesch* [2007] and *Fernandes et al.* [2007]) and August–October in the Arctic Ocean (*Boisvert and Stroeve* [2015])
			January–October in the Chukchi, Beaufort, Laptev, and East Siberian Seas (*Screen and Simmonds* [2010b])
		D	January–October in the Greenland Sea (*Screen and Simmonds* [2010b]), December–February in the Kara, Barents, and Greenland Seas (*Boisvert et al.* [2013])
Net precipitation	Annual	I	Northern continents, especially Eurasia (*Richter-Menge and Overland* [2009], *Haine et al.* [2015], and *Bring et al.* [2016]).

[a]An increasing trend is denoted by I and a decreasing one by D. References are made only to papers with the study period extending to 21st century. For the numerical values of the trends and the exact study periods, see the references cited.

particularly challenging to understand the nonlinear interactions of different processes, which are acting simultaneously on different spatial and temporal scales. This challenge in understanding is reflected in problems of modeling the processes.

3. Past Changes and Key Drivers

In the following subsections, we summarize changes in the Arctic atmospheric water cycle during the recent decades. Information on the changes is mostly based on in situ and remote sensing observations and reanalyses, but for many variables and regions the amount of observations is not adequate to make firm conclusions on trends (section 3.1). Information on the performance of climate models in historical simulations (section 3.2) is important in evaluating the reliability of future scenarios based on climate models. Further, historical climate model experiments have been a major approach to understand the drivers of the past changes (section 3.3). The modeling strategies are discussed in depth in *Lique et al.* [2016].

3.1. Observed Changes

Trends in key freshwater variables detected over the latest decades are summarized in Table 1 with respect to the season, region, and sign of the trend. Papers with the study period ending before year 2000 are not included. For the numerical values and exact periods of the trends, see the references cited in Table 1.

3.1.1. Atmospheric Moisture and Clouds

Widespread increases in annual near-surface specific humidity have occurred over the pan-Arctic region over the past several decades. The largest increases have occurred during summer and early autumn. Specific humidity has increased also during winter, spring, and all autumn, except for localized decreases (Table 1). Considering the annual mean vertically integrated water vapor (precipitable water), epoch differences between years 1986–2013 and 1958–1985, based on the Japanese 55-year reanalysis (JRA-55) [*Kobayashi et al.*, 2015], suggest increases everywhere in the circumpolar Arctic and sub-Arctic (Figure 3). We chose JRA-55 as it is the only recent reanalysis covering the entire period since 1958 with an advanced data assimilation method (four-dimensional variational data assimilation). Increases in precipitable water have prevailed also in summer

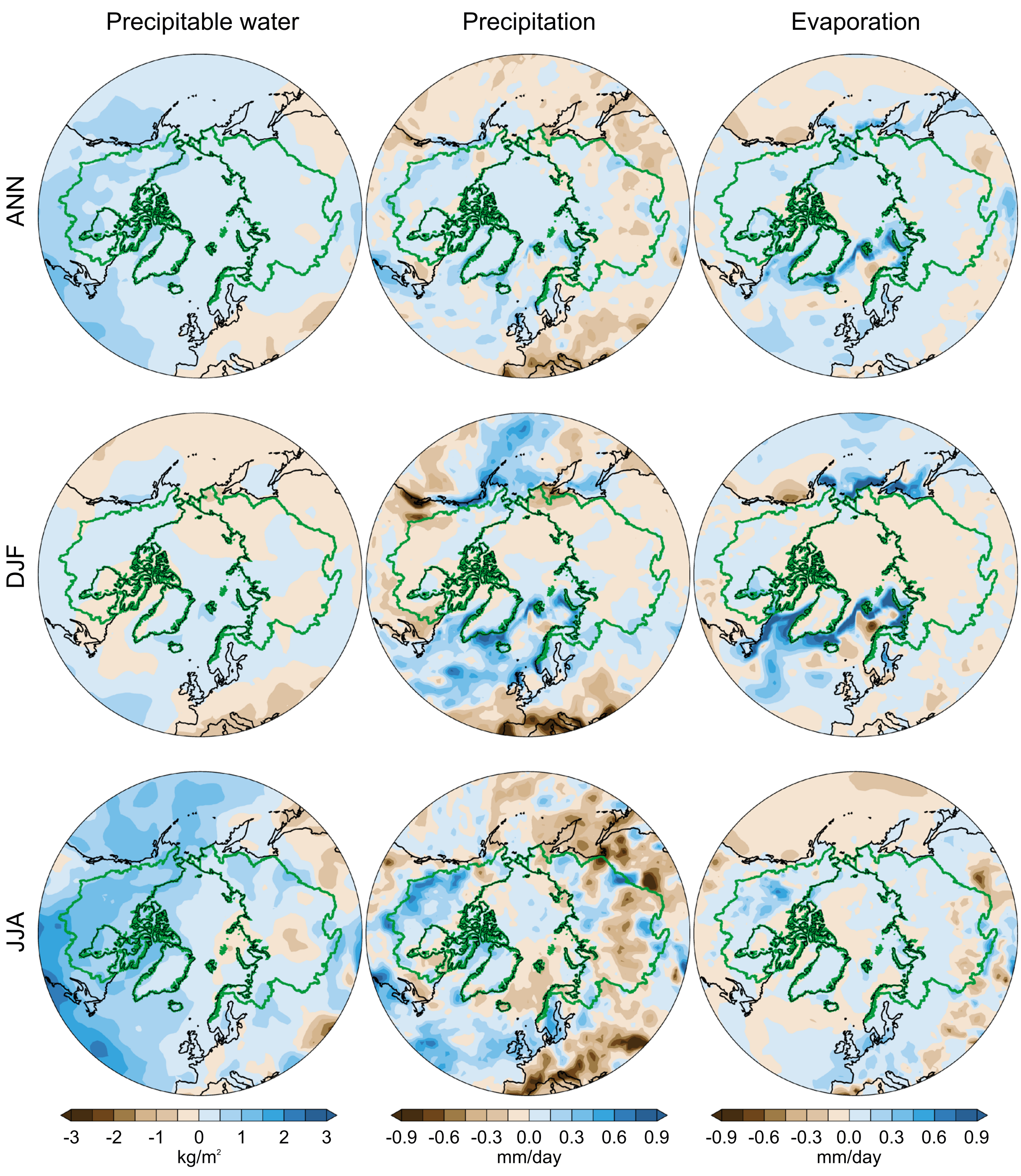

Figure 3. Epoch differences between 1986–2013 and 1958–1985 for precipitable water, precipitation, and evaporation on the basis of JRA-55 reanalysis for annual means, winter (DJF), and summer (JJA). The green lines indicate the boundaries of the Arctic river catchment.

(June-July-August, JJA), the main exceptions being decreases in land areas in western Siberia and in parts of the Kara and Barents Seas. Evaporation has increased in the same sea areas. Hence, the decrease in precipitable water can only be due to a decrease in moisture transport to these sea areas. Increases in precipitable water dominate also in winter (December-January-February, DJF), but the air column has become drier in the Labrador Sea, northeastern Canada, and eastern Siberia and surrounding seas (Figure 3). The seasonal and regional changes are, however, sensitive to the time period studied and reanalysis applied.

Quantitative interpretation of visual and satellite-based cloud data is not straightforward, but there are indications of increases in cloudiness over the Arctic, especially in low clouds during the warm season [*Walsh et al.*, 2011a, 2011b]. Annual zonally averaged midlatitude and high-latitude terrestrial cloud cover seems to have generally increased, except in the northernmost parts of Eurasia (Table 1). It should be noted that many studies have focused on the total cloud cover, but changes in clouds in various altitudes may be opposite. The increase in terrestrial cloud cover has been primarily due to an increase in low-level clouds [*Eastman and Warren*, 2013; *Nahtigalova*, 2013]. On the contrary, on the basis of both satellite observations and ERA-40 reanalysis, *Schweiger et al.* [2008] found that over the Arctic Ocean the sea ice decline has been related to a decrease in low-level cloud coverage and a simultaneous increase in midlevel clouds. Seasonal trends in total cloud cover have been mixed, the sign depending on the region (Table 1), exact study period, and possibly also on the data set used [*Wang and Key*, 2005; *Liu et al.*, 2008; *Eastman and Warren*, 2013; *Khlebnikova and Sall*, 2009; *Sun and Groisman*, 2000].

3.1.2. Precipitation and Evapotranspiration

Changes in precipitation and evapotranspiration over the last 56 years are illustrated in Figure 3 as epoch differences between 1986–2013 and 1958–1985. The results suggest a general increase in the annual mean precipitation and evaporation in the Arctic but with large spatial differences. The changes are also sensitive to the time period and reanalysis applied. Hence, the literature-based results reported in Table 1 suggest changes that partly differ from those shown in Figure 3. Annual precipitation in circumpolar midlatitude and high-latitude regions has increased over the latest decades (Table 1); however, considerable spatial and temporal variability exists [*McBean et al.*, 2005; *Min et al.*, 2008; *Hartmann et al.*, 2013]. In the Arctic, recent annual precipitation generally exceeds the mean of the 1950s by about 5% [*Walsh et al.*, 2011a, 2011b].

Winter precipitation trends have been mixed with increases and decreases reported for different regions (Table 1). Summer precipitation has generally increased across Canada, but significant trends with both signs have been observed in the upper Mackenzie River basin (Table 1). Spring precipitation has increased in Canada, while mixed trends have been detected for autumn. Summer and autumn precipitation have increased over East Siberia and decreased over central Eurasia (Table 1). Several different variables are used to quantify precipitation extremes. Significant trends have been detected in the number of days with heavy precipitation, daily precipitation intensity, and the number of consecutive dry days. In the first two variables, the sign of the trend has varied between regions, whereas the number of consecutive dry days has decreased in western Canada and western Eurasia (Table 1). Evaluation of precipitation trends is naturally hampered by measurement errors, which usually depend on the precipitation gauge type. Among the many error sources, wind is serious particularly in the case of snowfall. Over recent decades, however, corrections accounting for known precipitation gauge biases have been applied and, according to our evaluation, most of the discrepancies in precipitation trend estimates shown in Table 1 are caused by differences in the study regions and periods instead of measurement errors.

Changing patterns of winter precipitation and temperature have resulted in changes to the ratio of snow to total precipitation, with decreasing trends primarily in spring and autumn but increasing trends in winter in northern Canada and Siberia (Table 1). Increasing winter precipitation in the Ob and Yenisey basins has been accompanied by an increase in maximum snow depth and snow water equivalent [*Bulygina et al.*, 2011]. The decreasing trends have contributed to shortening of the snow cover duration [*Liston and Hiemstra*, 2011; *Zhang et al.*, 2011]. Further, summer snowfall has strongly decreased (40% between 1989 and 2009) over the Arctic Ocean [*Screen and Simmonds*, 2012], mostly due to a change from snowfall to rain in a warming climate.

Rising temperatures, particularly in high latitudes, and changing patterns of precipitation have been accompanied by increases in annual evapotranspiration over most of Canada (Table 1). Over sea areas, information on historical changes in evaporation is mostly based on reanalyses, hence being less reliable than information from land areas. According to ERA-Interim, evaporation in January through October has had an increasing trend from 1989 to 2009 in the Chukchi, Beaufort, Laptev, and East Siberian seas but a decreasing trend in the Greenland Sea [*Screen and Simmonds*, 2010b]. Recent methods to estimate evaporation applying air moisture based on Atmospheric Infrared Sounder (AIRS) satellite data and wind speed based on ERA-Interim have yielded scattered results depending on the version of AIRS data applied [*Boisvert et al.*, 2013; *Boisvert and Stroeve*, 2015], with the latest version suggesting an increasing evaporation trend in August–October in 2003–2013 in the Arctic Ocean [*Boisvert and Stroeve*, 2015]. The epoch differences (1986–2013 compared to 1958–1985) based on JRA-55 suggest an increased annual mean evapotranspiration over both land and sea areas in most of the circumpolar

Arctic and sub-Arctic (Figure 3). In addition to evapotranspiration, transport from lower latitudes is an important source for atmospheric moisture in the Arctic. See section 3.3 for changes in the transport.

Net precipitation has increased over northern continents, especially Eurasia (Table 1), seen as an increase in the river discharge to the Arctic Ocean. The trends, however, depend on the region and period addressed. Considering the JRA-55-based epoch differences between 1986–2013 and 1958–1985, in a large part of Eurasian midlatitudes as well as in smaller regions in Canada and USA, the annual mean precipitation has decreased but evapotranspiration has simultaneously increased (Figure 3). Accordingly, these predominantly midlatitude regions have changed toward "water poor." An increase in precipitation with a simultaneous decrease in evapotranspiration has taken place further north, e.g., in the Baffin Island and coasts of the Kara Sea, which have become more "water rich" (Figure 3). Over longer time periods, since 1930s, the general trend in the Arctic river catchments in Eurasia has been an increasing discharge, i.e., toward more water rich [*Richter-Menge and Overland*, 2009].

Analyses on historical changes in net precipitation over the Arctic Ocean have been limited. According to *Rawlins et al*. [2010], net precipitation over the Arctic Ocean shows no trend for the period 1979–2007, estimated on the basis of the JRA-25 reanalysis. According to *Haine et al*. [2015], however, net precipitation over the Arctic Ocean has increased from 2000 ± 200 mm yr^{-1} in 1980–2000 (mostly based on *Serreze et al*. [2006]) to 2200 ± 220 mm yr^{-1} in 2000–2010 (Figure 1). See *Carmack et al.* [2016] for more discussion on precipitation and evaporation over Arctic seas.

3.2. Historical Simulations

3.2.1. Clouds

Analysis of historical simulations on Arctic clouds has focused mainly on their climatology, aiming to understand associated feedback process contributing to Arctic climate change. The Advanced Very High Resolution Radiometer Polar Pathfinder (APP-x) data set [e.g., *Wang and Key*, 2003] has been widely used in the analysis, showing maximum climatological cloud cover over the Norwegian and Barents Seas. The cloud cover data also demonstrate an obvious seasonal cycle, with peak values occurring in summer. Comparing with these observed climatological features, the ensemble mean of Climate Model Intercomparison Project phase 3 (CMIP3) models realistically captures average aspects of the amount and spatial distribution of Arctic clouds, as well as predominant types of low clouds, during summer and fall, but exhibits a large across-model spread occurring during winter and spring [*Vavrus et al*., 2009; *Karlsson and Svensson*, 2011]. A continuing analysis indicates that the large across-model spread remains in the follow-up CMIP5 model historical simulations, suggesting no obvious improvement has been made [*Karlsson and Svensson*, 2013].

Cesana and Chepfer [2012] compared the cloud vertical structure in five of the Atmospheric Model Intercomparison Project models against the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation satellite measurements. The results indicated that, during 1979–2008, the models performed rather well for the low cloud amount (ranged from 37% to 57% between models, compared to the observed 44%) but did not capture the observed seasonal cycle. Examination of a stand-alone Community Atmospheric Model, version 5 (CAM5) simulation exhibited insufficient amount of cloud cover throughout the year [*English et al*., 2014]. The across-model spread in cloud variables contributes to the across-model spread in surface radiative forcing, impacting cloud albedo effects and leading to possibly biased understanding and assessment of Arctic climate change [*Karlsson and Svensson*, 2013]. In addition, it is also argued that cloud cover is not a well-defined variable, in general and in particular for the Arctic, as substantial portions of Arctic cover are optically thin [*Koenigk et al*., 2013]. For more discussion on challenges in historical simulations on Arctic clouds, see *Lique et al.* [2016].

3.2.2. Precipitation and Evapotranspiration

The historical climatology and changes in precipitation and net precipitation in the Arctic have been simulated by various state-of-the-art climate models [*Walsh et al*., 1998; *Holland et al*., 2006; *Kattsov et al*., 2007; *Bengtsson et al*., 2011]. Observed climatological precipitation has a distinct seasonal cycle, peaking in summer. It is generally high on the North Atlantic side of the Arctic, in particular over the Greenland and Norwegian Seas, and with an extension to the Eurasian shelf seas. *Kattsov et al*. [2007] analyzed simulation results from the 21 models participating in the CMIP3, focusing on comparison of the simulated precipitation and net precipitation against observational or observationally based data sets over the Arctic (defined as north of 70°N and the four largest Eurasian and American Arctic river basins). Although the observations are uncertain (section 3.1.2) and there are large areas with no observations, *Kattsov et al*. [2007] concluded that the multimodel ensemble mean overestimates precipitation over the shelf seas and adjacent lands and underestimates in the central Arctic Ocean during the winter season. When compared with the observationally based estimates of precipitations from 1960 to 1989 by *Bryazgin* [1976] and *Khrol* [1996], and with ERA-40, the CMIP3 simulations show spatially dependent biases with an overestimate over the western Arctic and an underestimate over the eastern Arctic [*Kattsov et al*., 2007]. Overall, the multimodel simulated, regionally averaged, annual mean precipitation appears realistic, though substantial differences occur across individual model simulations, with the largest spread in early fall.

Positive trends of annual mean precipitation have been detected at northern high latitudes [*Trenberth*, 2011]. The ensemble mean of the CMIP3 models, or simulations by individual models, such as Community Climate System Model version 3 (CCSM3) and fifth-generation European Centre/Hamburg model (ECHAM5), generally captures this long-term change over the Arctic Ocean in the historical simulations, consistent with the results from an earlier generation of climate models [*Kattsov and Walsh*, 2002; *Holland et al*., 2006; *Kattsov et al*., 2007; *Bengtsson et al*., 2011]. The modeled trend in annual mean precipitation is mainly due to positive changes in the winter season. The model simulations show an increasing trend of evapotranspiration over the pan-Arctic terrestrial region [*Kattsov et al*., 2007; *Rawlins et al*., 2010]. However, large diversity across models exists. It is most challenging to examine model's performance due to a lack of enough observations.

3.3. Drivers

Human-induced global warming is a prominent climate feature during past decades. This warming trend is considered as a fundamental driver for the observed long-term changes in water cycle during this time period. Theoretical estimates based on the Clausius-Clapeyron relation and observations have indicated an increase in lower tropospheric moisture content (specific humidity) at a rate of about 7% K^{-1} [*Wentz and Schabel*, 2000; *Held and Soden*, 2006]. This moistening trend could be amplified over the Arctic due to enhancement of poleward moisture transport [*Zhang et al*., 2013]. In fact, the atmospheric circulation has experienced shifts with more meridionally oriented wind flow since the late 1990s, strengthening the poleward moisture transport [*Zhang et al*., 2008]. This has increased the atmospheric moisture content over the largest Eurasian river basins [*Zhang et al*., 2013]; the consequences are discussed in *Bring et al*. [2016].

In addition to human-induced warming, drivers of changes in the Arctic atmospheric water cycle are related to inherent variability of the polar jet stream, storm tracks, and related synoptic-scale weather. Part of the variability in the jet stream and storm tracks can be characterized by large-scale circulation modes such as the Arctic Oscillation (AO), North Atlantic Oscillation (NAO), East Atlantic Pattern [*Woollings and Blackburn*, 2012], Pacific Decadal Oscillation, and Southern Oscillation [*Newton et al*., 2014a, 2014b], but the relationships vary regionally [*Overland et al*., 2015]. When the NAO/AO index is positive, the storm tracks shift northward and winters are typically mild in northern Eurasia and the eastern United States but cold in Greenland, Canada, Alaska, and eastern Siberia. When the NAO/AO index is negative, the storm tracks shift southward and temperature field is reversed [*Cohen et al*., 2014]. Several studies suggest that the number of cyclones entering the Arctic has increased during recent decades [*Zhang et al*., 2004; *Trigo*, 2003; *Sorteberg and Walsh*, 2008; *Sepp and Jaagus*, 2011], favoring an increase in the moisture transport from lower latitudes, but according to *Sepp and Jaagus* [2011] the number of cyclones formed north of 68°N has not increased. Storms have a strong effect on the freshwater cycle, e.g., via ocean mixing [*Carmack et al*., 2016]. Storminess (the occurrence of storms) in the Arctic has varied in decadal time scales [*Atkinson*, 2005; *Mesquita et al*., 2010] and increased in some locations in the North American Arctic, but there are no indications of systematic increases in storminess in the circumpolar Arctic since 1960s [*Walsh et al*., 2011a, 2011b]. Note that storminess differs from cyclone activity, as most Arctic cyclones are not associated with storm-level winds. Due to decadal changes in the availability of in situ and remote sensing data, however, evaluation of past changes in Arctic cyclone activity and storminess is difficult [*Döscher et al*., 2014].

Not only cyclones but also anticyclones are essential drivers of the Arctic freshwater cycle. Of particular importance is the Beaufort High [*Serreze and Barrett*, 2011], centered over the Beaufort Sea, which is typically strong when the AO is negative [*Rigor et al*., 2002] and has strengthened in summer during the most recent decades [*Wu et al*., 2014]. A strong Beaufort High favors clear skies [*Kay et al*., 2008] and reduced precipitation, but otherwise the *direct* effects of the Beaufort High on the atmospheric water cycle have not received much attention. As a driver of the Beaufort Gyre, the Beaufort High is, however, very important for the oceanic freshwater budget [*Haine et al*., 2015; *Carmack et al.*, 2016] and variability in the strength and location of the Beaufort High has played a role in the rapid loss of Arctic sea ice [*Rigor and Wallace*, 2004; *Ogi and Wallace*, 2007; *Wang et al*., 2009].

As mild winters in high latitudes are associated with increased atmospheric moisture, a positive relationship has been generally identified between the total cloud cover, Arctic Oscillation index, and a poleward shift of synoptic-scale storm tracks [*Eastman and Warren*, 2010; *Serreze et al*., 1993; *Zhang et al*., 2004]. However, low-level clouds seem less correlated with the large-scale atmospheric circulation or synoptic-scale weather regimes [*Shupe et al*., 2011]. An analysis of CMIP3 model simulations suggests that local evaporation would be a leading candidate for the Arctic cloud response due to their high correlation [*Vavrus et al*., 2009]. In addition to air temperatures and moisture variables, variations in the jet stream and storm tracks have strong effects on the ocean [*Carmack et al.*, 2016].

Drivers for the observed changes in Arctic precipitation and evapotranspiration may also be inferred from climate model simulations under the observed forcing in the twentieth century. Increases in precipitation or precipitation minus evaporation are identifiable in present-day climate simulations [*Holland et al*., 2006; *Kattsov et al*., 2007; *Bengtsson et al*., 2011; *Koenigk et al*., 2013; *Bintanja and Selten*, 2014], suggesting the forcing role of the anthropogenic effects and resulting Arctic sea ice retreat. However, evaporation exhibits an inconsistent response across models. For example, it shows a decrease in ECHAM5 [*Bengtsson et al*., 2011] and an increase in CCSM3 [*Holland et al*., 2006] and CMIP5 models [*Bintanja and Selten*, 2014]. Evaluation of modeled evaporation is also subject to a lack of direct observations, in particular over the Arctic region. Nevertheless, according to the increase in the net poleward moisture transport detected from reanalysis data [*Zhang et al*., 2013] and simulated by climate models [*Bengtsson et al*., 2011], the rate of evaporation increase must be less than that of precipitation increase.

To attribute the long-term increasing trend of annual mean precipitation, *Min et al*. [2008] employed a standard optimal detection approach to compare observed and simulated Arctic land precipitation. They found that the increase in anthropogenic forcing has dominated the observed upward trend of high-latitude precipitation in the second half of the twentieth century (Figure 4). Meanwhile, they also indicated that the model simulated precipitation response to the anthropogenic forcing is weaker than that in the observations, which is consistent with the findings in other studies [*Holland et al*., 2006; *Kattsov et al*., 2007; *Bengtsson et al*., 2011; *Koenigk et al*., 2013; *Lique et al*., 2016]. Note that AO has generally been used to represent intrinsic atmospheric variability in detection and attribution analysis. However, anthropogenic forcing may also affect AO, which further affects atmospheric moisture transport from lower latitudes, precipitation, and river discharge into the Arctic Ocean [*Zhang et al*., 2008, 2013]. But the interactions between intrinsic atmospheric variability and anthropogenically forced changes in the atmospheric circulation have not been well simulated and understood. These problems may have contributed to the weaker precipitation in models than in observations.

3.4. Summary

Over longer periods and large areas, there seems to be a trend toward wetter conditions, but when smaller regions and shorter time periods are analyzed, variable trends are detected. Models capture the overall wetting trend but have problems in reproducing the regional details [*Lique et al.*, 2016]. The overall increase in precipitation is linked to the general warming trend, partly driven by anthropogenic forcing, and amplified in the Arctic via various positive feedbacks.

4. Projected Changes and Key Drivers

In the following subsections, we briefly summarize atmospheric freshwater projections based on model simulations performed as part of CMIP5. More information on the various challenges, error sources, and strategic aspects in modeling of the future freshwater in the Arctic are discussed in *Lique et al.* [2016]. Here all statements are based on the CMIP5 multimodel mean (or median), rather than individual models, unless stated otherwise. The CMIP5 projections are broadly consistent with those from the CMIP3 models analyzed by *Kattsov et al*. [2007]. We focus on multidecadal projections by the end of the 21st century, although near-term projections (up to 2050) are broadly similar in sign, but with reduced magnitude and greater uncertainty due

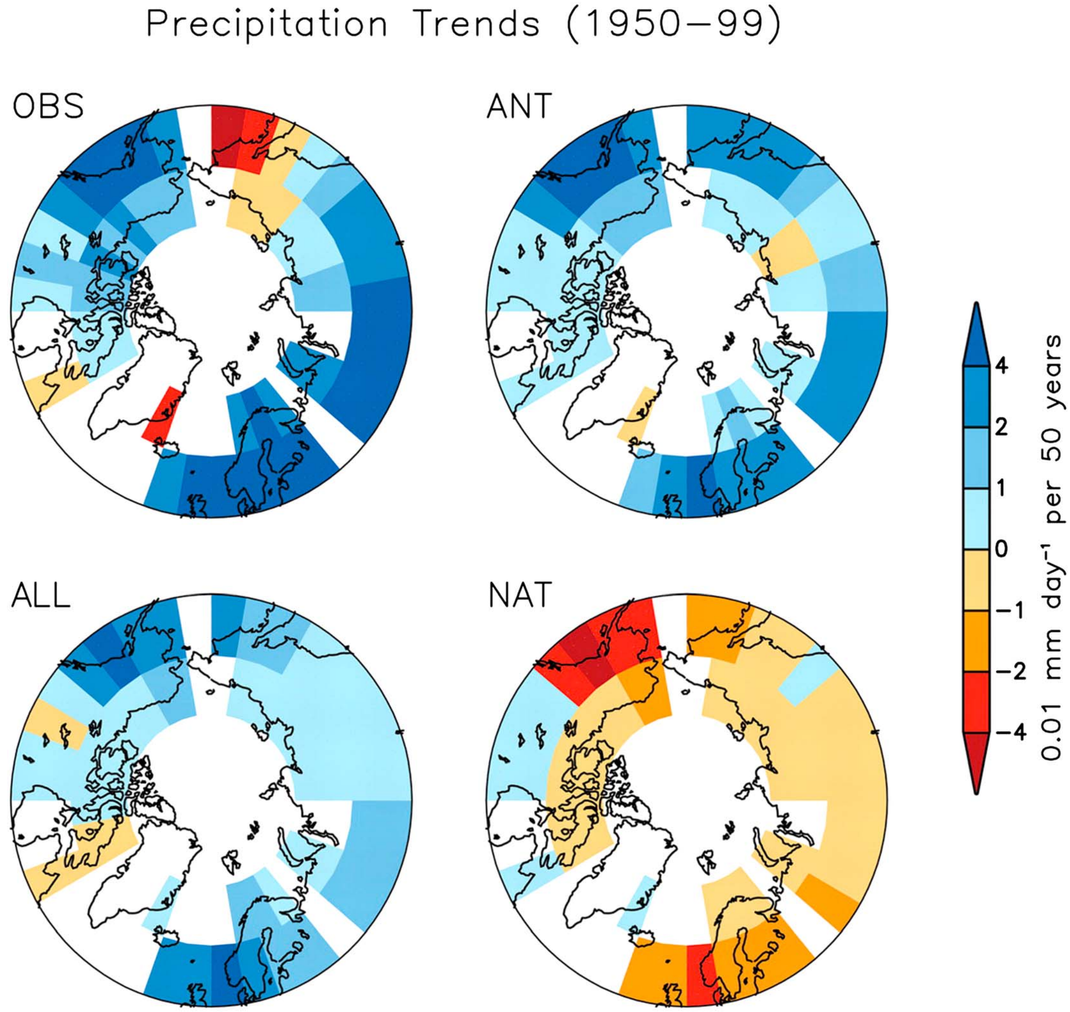

Figure 4. Large-scale precipitation trends (0.01 mm d^{-1} per 50 years) during 1950 to 1999 based on observations (OBS) and CMIP3 model simulations driven by different forcings: anthropogenic (ANT; greenhouse gases and sulfate aerosols), natural (NAT; solar and volcanic), and both of them (ALL). Areas with less than 40 years of observations are marked with white space. Reproduced with permission from *Min et al.* [2008].

to natural variability. Beyond 2050, the projected changes are generally of the same sign in all emission scenarios but are larger in magnitude for higher greenhouse gas concentration pathways. To first order, the magnitude of projected changes in the atmospheric freshwater system scales with projected temperature changes.

4.1. Model Projections

4.1.1. Humidity

Specific humidity (SH) is projected to increase with warming, according to the Clausius-Clapeyron relation. Simulated SH increases in the Arctic are sustained by increased local evaporation and enhanced moisture flux convergence [*Skific and Francis*, 2013]. Since warming is amplified at the surface, SH rises are also expected to be largest in the lowermost atmosphere. The CMIP5 model ensemble suggests that relative humidity (RH) will decrease [*Collins et al*., 2013]. Over northern land regions, RH is projected to decrease in the annual, winter, and summer mean. This decrease arises because land regions warm faster than open ocean regions. When maritime air masses are advected over land, the air is warmed and its relative humidity drops, as any further moistening of the air over land is insufficient to maintain constant RH [*Collins et al*., 2013]. Over the Arctic Ocean, the near-surface RH is very sensitive to lead occurrence, which depends on sea ice dynamics over a broad range of spatial scales. Climate models can hardly simulate it reliably enough to allow solid conclusions of future evolution of RH.

4.1.2. Clouds

The CMIP5 models project a general increase in annual mean total cloud fraction by a few percent over the Arctic Ocean and high-latitude land regions by the late 21st century [*Collins et al.*, 2013]. Further south in the midlatitudes, cloud cover is projected to decrease consistent with reduced RH. However, these projected changes are not robust across the CMIP5 models, reflecting model differences in cloud physics and cloud parameterizations. These discrepancies appear to arise primarily due to differences in simulated low clouds [*Collins et al.*, 2013]. In the Arctic, the cloud discrepancies are seasonally largest in autumn. For example, while one CMIP5 model (EC-Earth) projects a decrease in autumn cloud cover [*Koenigk et al.*, 2013], another (CCSM4) model projects a large increase [*Vavrus et al.*, 2012]. In both these models, the largest cloud cover changes occur in the months and regions of greatest Arctic sea ice loss, but the responses are of opposite sign. Thus, the cloud cover response to reduced sea ice cover is a key area of uncertainty in projected cloud cover trends.

4.1.3. Precipitation

The CMIP5 models project a robust increase in precipitation (*P*) over the Arctic and midlatitudes, related to warming and increased available moisture [*Collins et al.*, 2013; *Laine et al.*, 2014; see also *Lique et al.*, 2016] (Figure 5). A robust increase in precipitation with scattered results for trends in cloudiness (section 4.1.2) and seasonally and regionally varying trends for relative humidity (section 4.1.1) is an interesting combination. It can probably be explained by the fact that if the clouds contain more water, precipitation may increase even in regions where relative humidity and cloud coverage decrease, i.e., precipitation intensity increases. In the winter half-year (October–March), Arctic mean *P* is projected to increase by 35% and 60%, under medium and high greenhouse gas representative concentration pathways (RCP), respectively (RCP4.5 and 8.5), relative to the period 1986–2005 [*Intergovernmental Panel on Climate Change* (*IPCC*), 2013]. The projected *P* changes in the warmer months (April–September) are weaker: 15% and 30%. Spatially, the largest relative changes are projected over the Arctic Ocean [*IPCC*, 2013; *Laine et al.*, 2014].

An annual mean decrease in snowfall is projected over North America, Northern Europe, and midlatitude Asia [*Krasting et al.*, 2013]. These decreases are driven by reductions in the snowfall-to-precipitation ratio (i.e., a shift from snow to rain), which is temperature dependent (see section 5.4 and *Wrona et al.* [2016], on ecological impacts). Seasonally, the changes are largest in spring and autumn, associated with an earlier snow-rain transition in spring and later rain-snow transition in autumn [*Krasting et al.*, 2013]. In high latitudes (e.g., northern Siberia) and at high altitudes (e.g., central Greenland), where air temperatures remain sufficiently cold, snowfall is projected to increase, especially in winter.

In addition to projected increases in mean *P*, there is increasing evidence that *P* extremes will also increase in frequency and/or intensity. According to *Kharin et al.* [2013], relative changes in the intensity of precipitation extremes generally exceed relative changes in annual mean precipitation. On the basis of eight CMIP5 models, *Toreti et al.* [2013] found out that the 1 in 50 year return level of daily precipitation is projected to increase in all seasons over the middle and high latitudes, being reliable and consistent between models particularly over northern Eurasia in winter and the North Pacific and northwestern Atlantic/Arctic Ocean in summer. Due to the Arctic amplification (section 2.1), in consistence with the Clausius-Clapeyron constraint, the largest increases are projected poleward of 60°N. The 20 year return level of annual extremes of daily precipitation is also projected to increase over the Arctic and Arctic River catchments, by around 2–6% per degree of local warming [*Kharin et al.*, 2013]. Other indices of extreme *P*—namely total wet day precipitation, very wet day precipitation, maximum 5 day precipitation, and heavy precipitation days—are all projected to increase over the midlatitude to high-latitude northern landmasses, especially north of 60°N [*Sillmann et al.*, 2013]. Increases of 20–30% in the maximum 5 day precipitation are projected for Alaska, northern and eastern Canada, Greenland, and northern Eurasia. The increases are less pronounced in summer than in winter with the strongest increases in northern Asia of about 40% in winter and 20% in summer [*Sillmann et al.*, 2013]. In northern Europe, a CMIP5 ensemble shows a consistent increase only for winter. In Alaska, northern and eastern Canada, Greenland, and northern Asia these increases in precipitation are accompanied by a decrease in the number of consecutive dry days, whereas in northern Europe no consistent change is projected for dry days [*Sillmann et al.*, 2013].

4.1.4. Evaporation

Evaporation (*E*) is projected to increase in winter over the Arctic and northern continents, except for in the North Atlantic region and Greenland [*Laine et al.*, 2014; see also *Lique et al.*, 2016] (Figure 5). These increases are consistent with increases in humidity and *P*, and overall warming that increases potential *E*. The projected decrease over the North Atlantic is related to a reduced air-sea humidity contrast. In summer, *E* is projected to decrease

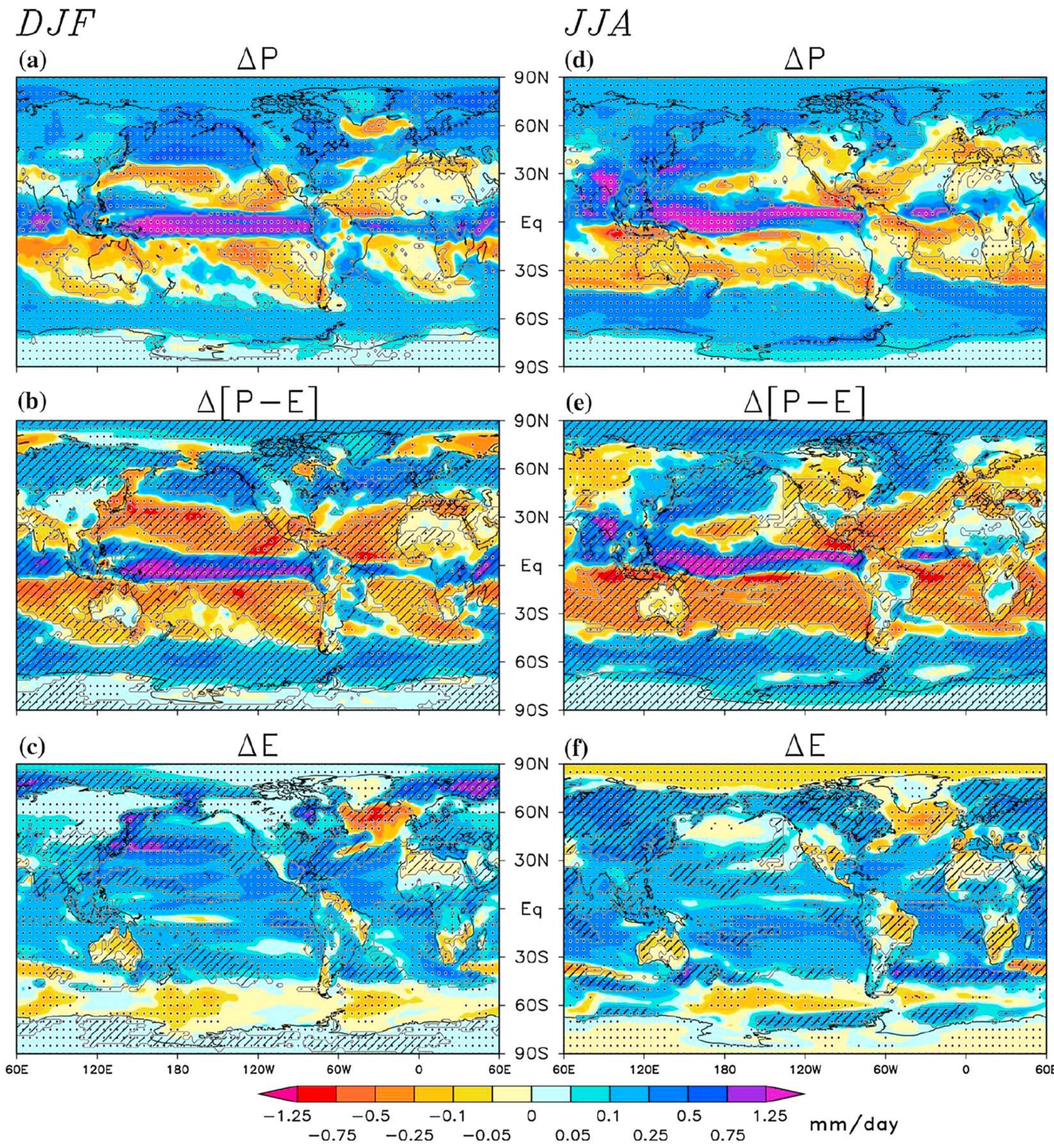

Figure 5. (a and d) Multimodel mean changes in precipitation (*P*), (b and e) precipitation − evaporation (*P* − *E*), and (c and f) evaporation (*E*) for the winter (DJF; Figures 5a–5c) and summer (JJA; Figures 5d–5f) seasons. Regions are *dotted* in which the sign of the changes is robust among the models at the 95% confidence level. In the Figures 5b and 5e and Figures 5c–5f, regions are hatched in which the particular contribution dominates the precipitation changes indicated in Figures 5a and 5d. Reproduced with permission from *Laine et al.* [2014].

over the Arctic Ocean and subpolar ocean [*Laine et al.*, 2014] (Figure 5). This is because near-surface air temperature and humidity rise, reducing the air-sea temperature and moisture contrasts and therefore, reducing *E*. Over the northern continents, summer *E* increases in line with warming.

4.1.5. Precipitation − Evaporation

Wintertime *P* − *E* is projected to increase over the central Arctic Ocean but decrease over the coastal seas where the winter sea ice concentration is reduced, allowing more evaporation [*Laine et al.*, 2014; see also *Lique et al.*, 2016] (Figure 5). The projected sea ice loss leads to large local increases in *E* that exceed the *P* increases. Over the central Arctic Ocean, where winter sea ice cover remains limiting air-sea exchange, the warming-driven *P* increases exceed the *E* increases. Over northern landmasses *P* − *E* increases during winter, as *P* increases faster than *E* in response to warming. Conversely in summer, *P* − *E* decreases over northern continents due to larger increases in *E* than *P* [*Laine et al.*, 2014] (Figure 7). Over the Arctic Ocean and subpolar oceans, summer *P* − *E* increases in response to increased *P* and decreased *E*. For the high latitudes, the physical consistency and the similar behavior across multiple generations of models and forcing scenarios indicate that the projected *P*, *E*, and $P - E$ changes discussed above are likely with high confidence [*Collins et al*., 2013].

4.2. Drivers of Projected Change

4.2.1. Surface Moisture Budget

An analysis of the CMIP5 models found that over 50% of the projected Arctic precipitation increase by the late 21st century is related to local evaporation changes, closely tied to the retreat of sea ice [*Bintanja and Selten*, 2014] (Figure 6). Seasonally, the contribution due to local evaporation is largest in winter, and the contribution due to moisture transport is largest in late summer and autumn. Several studies have sought to isolate the influence of projected Arctic sea ice loss by running atmospheric model simulations prescribed with future sea ice conditions but unchanged sea surface temperatures or radiative forcing. Such simulations project an intensified Arctic hydrological cycle in response to future sea ice loss, with the largest changes near regions of sea ice loss and in the cold season [*Deser et al*., 2010; *Peings and Magnusdottir*, 2014; *Deser et al*., 2014]. The sea ice retreat will lead to more open water and for more of the year, giving rise to new near-coastal hydrological regimes with implications for marine and freshwater ecosystems [*Carmack et al*., 2016; *Wrona et al*., 2016]. Over land, projected reductions in snow cover enhance lower tropospheric warming [*Alexander et al*., 2010; *Henderson et al*., 2013]. However, the impact of projected snow cover declines on the atmospheric moisture budget has not been explored.

4.2.2. Poleward Moisture Transport

Atmospheric transport of moisture into the Arctic region is projected to increase in response to rising greenhouse gas (GHG) concentrations. This increase can be partially explained by simple thermodynamics (*Held and Soden* [2006] and see also discussion in *Lique et al.* [2016]. As the atmosphere warms, it can hold more moisture. Humidity increases more rapidly at lower latitudes for a given temperature increase, leading to a larger poleward moisture gradient. Both increased atmospheric moisture content and a larger poleward moisture gradient lead to a larger transport of poleward moisture into the Arctic. Approximately 75–80% of the total projected annual change in moisture transport across 70°N, between the late 20th and 21st centuries, is thermodynamically driven, i.e., due to change in the meridional gradient of specific humidity [*Skific et al*., 2009; *Skific and Francis*, 2013]. Although smaller, the dynamic term is also positive and related to an increase in low-pressure systems over the central Arctic that transport substantial moisture into the Arctic [*Skific et al*., 2009; *Skific and Francis*, 2013].

North Atlantic atmospheric rivers are projected to become stronger and more frequent [*Lavers et al*., 2013], transporting increased moisture into the North Atlantic and Eurasian sectors of the Arctic, which via increasing precipitation contributes to freshening of the seas [*Carmack et al*., 2016]. This projected increase is predominantly a thermodynamic response to warming and moistening [*Lavers et al*., 2013], rather than due to dynamical changes in storms and their associated frontal features.

Haine et al. [2015] reviewed climate model projections for the end of the 21st century and concluded that the atmospheric moisture flux convergence over the Arctic Ocean and Canadian Arctic archipelago will be approximately equally large as the oceanic freshwater transport through the Bering Strait (~2500 km^3/yr, the oceanic transport calculated with respect to a reference salinity of 34.80 in the practical salinity scale), whereas the river runoff will be roughly double as large (5500 km^3/yr). These inflows will be balanced by the oceanic outflow, mostly via the Fram Strait and Davis Strait.

4.2.3. Storm Tracks

Projected upper tropospheric warming is larger in the tropics than in the extratropics, leading to an enhanced poleward temperature gradient at upper levels. This is expected to induce a poleward shift of the jet streams and accompanying storm tracks in middle latitudes [*Held*, 1993]. However, at the same time, projected Arctic sea ice loss greatly amplifies lower tropospheric warming in the Arctic relative to the midlatitudes. This low-level warming produces a weakening of the equator-to-pole near-surface temperature gradient, opposite to that at higher altitudes. A reduced lower tropospheric temperature gradient is associated with an equatorward shifted jet stream and storm tracks [*Hoskins and Woollings*, 2015; *Hall et al*., 2015]. The net effect of these two opposing effects in winter is close to cancelation, but in some models one influence is larger than the other. This results in weak and nonrobust winter storm track shifts across the CMIP5 models [*Barnes and Polvani*, 2013; *Harvey et al*., 2014]. Efforts to understand model spread in the atmospheric circulation responses to a rise in GHG have highlighted the competing effects of tropical upper

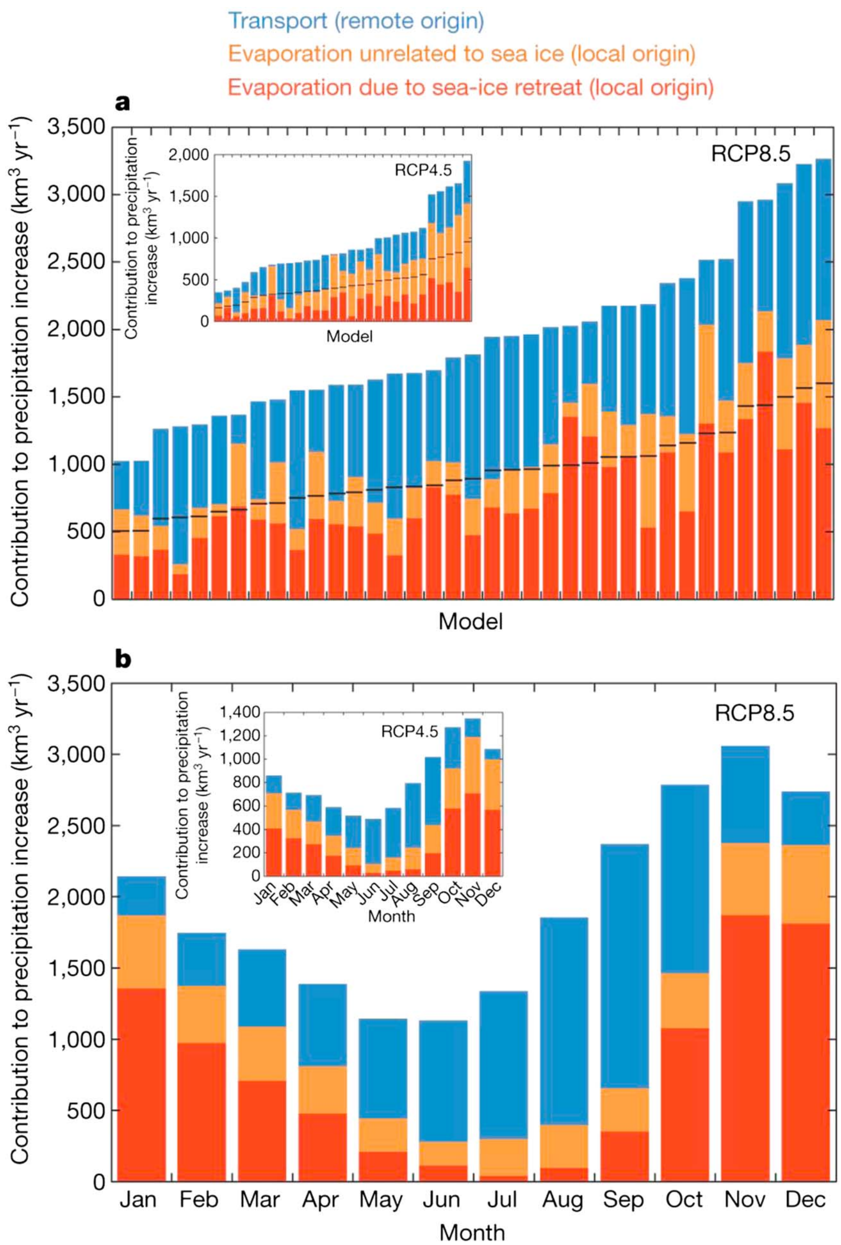

Figure 6. Simulated annual and monthly 21st century changes in Arctic mean precipitation, poleward moisture transport across 70°N (remote origin), and surface evaporation components (local origin). (a) Annual mean changes, where each bar denotes one CMIP5 model sorted according to the magnitude of the simulated precipitation change (horizontal black lines represent 50% of each column). (b) Seasonal changes, where each bar represents the monthly and multimodel mean. Changes are given for the strong (RCP8.5) and intermediate (insets, RCP4.5) forcing scenarios. The total Arctic surface evaporation (the sum of the red and orange bars) is separated into its ice retreat (red) and non-ice retreat (orange) components. The attribution of sea ice retreat to the total surface evaporation was evaluated by calculating monthly evaporative flux differences over only the ice retreat regions, which were then summed over the respective 10 year periods. Reproduced with permission from *Bintanja and Selten* [2014].

tropospheric warming and Arctic sea ice loss [*Cattiaux and Cassou*, 2013; *Haarsma et al.*, 2013; *Harvey et al.*, 2014; *Barnes and Polvani*, 2015; *Deser et al.*, 2015]. Stratospheric polar temperature trends have also been implicated as a source of spread in future tropospheric circulation trends [*Manzini et al.*, 2014]. In other seasons, the impact of Arctic sea ice loss on the atmosphere is weaker than in winter [e.g., *Deser et al.*, 2010], resulting in a robust poleward shift of the storm track in the warm part of the year [*Grise and Polvani*, 2014; *Barnes and Polvani*, 2015; *Deser et al.*, 2015].

4.3. Summary

The latest state-of-the-art climate model projections suggest a robust future intensification of the Arctic hydrological cycle. Mean precipitation and daily precipitation extremes are projected to increase over midlatitude and high latitude, largely in response to warming-driven increases in the moisture holding capacity of the air. The relative increases precipitation extremes are expected to exceed relative increases in mean values. This increased moisture comes from increased evaporation (itself largely due to the loss of sea ice cover) and enhanced poleward moisture transport. There is however, significant model divergence in future cloud cover trends, especially those arising from sea ice loss, and in the wintertime storm track changes.

5. Cross-System Effects

A schematic summary of cross-system impacts is presented in Figure 7, with focus on the effects of intensification of the Arctic atmospheric water cycle. The discussion is closely tied with the papers by *Bring et al*. [2016], *Carmack et al.* [2016], *Instanes et al.* [2016], and *Wrona et al*. [2016].

5.1. Remote Atmospheric Impacts

So far we have focussed attention on changes in the Arctic. However, the large changes in the Arctic freshwater storages may have implications for regions beyond the Arctic. Numerous recent studies have addressed the effects of Arctic sea ice loss and changes in Siberian snow cover on midlatitude weather and climate (see *Vihma* [2014] and *Cohen et al*. [2014] for reviews). Several linkages have been found, but many of them are regional and episodic and result from a combination of internal variability, lower tropospheric temperature anomalies [*Screen*, 2014], and midlatitude teleconnections [*Overland et al*., 2015]. There are physical mechanisms, such as the weakening of midtropospheric westerly winds due to the Arctic amplification, via which changes in the Arctic can influence midlatitude weather, but it is often difficult or impossible to prove if changes in the Arctic have been the main causal factor for observed anomalies in midlatitude weather [*Barnes and Screen*, 2015]. Some studies have suggested that the Arctic amplification of climate warming favors a weaker and more meandering jet stream, and this may have contributed to the extreme snowfall events in recent winters in the U. S. East Coast [e.g., *Francis and Skific*, 2015]. Also, model simulations by *Screen* [2013] suggested that Arctic sea ice loss induces a southward shift of the summer jet stream over Europe and increased northern European precipitation. This has probably contributed to the observed six consecutive wet summers from 2007 to 2012 in northern Europe.

Arctic amplification is robustly projected to continue in the future in response to increasing GHG concentrations [*IPCC*, 2013], further reducing the near-surface north-south atmospheric temperature gradient. Recent work has shown that future midlatitude climate change is moderated by the magnitude of projected Arctic amplification. For example, models that simulate larger Arctic amplification depict smaller poleward shifts of the winter jet streams and storm tracks [*Barnes and Polvani*, 2015; *Harvey et al*., 2014], with implications for precipitation over the Arctic river catchments [*Bring et al*., 2016].

5.2. Oceanic Impacts

Precipitation minus evaporation changes have direct impacts on the salinity of the Arctic Ocean above the halocline [*Carmack et al*., 2016, Figure 7]. However, the freshening effect of increased $P-E$ is small compared to freshwater input from increased river discharge (see section 5.3) as well as melting sea ice, glaciers, and ice sheets. Continents and islands play an important role in generating orographic precipitation; part of the precipitated water ends up in the Arctic and sub-Arctic Seas via river and meltwater discharge, iceberg calving, and glacier flow into the sea. Another important atmospheric influence on the ocean comes from wind forcing (Figure 7); changes in large-scale atmospheric circulation and storm tracks are summarized in section 3.3. A recent freshening of the western Arctic Ocean has been linked to a wind-driven spin-up of the Beaufort Gyre [*Giles et al*., 2012; *Carmack et al*., 2016], which is due to strengthening of the Beaufort High (section 3.3). Wind forcing is also vital in determining freshwater export via the Fram Strait [*Kwok et al*., 2013]. Considering sea ice export, there is no increasing trend [*Spreen et al*., 2009]. A small increase in the atmospheric pressure gradient across the Fram Strait would favor it, but the effect is compensated by a decrease in sea ice concentration [*Kwok*

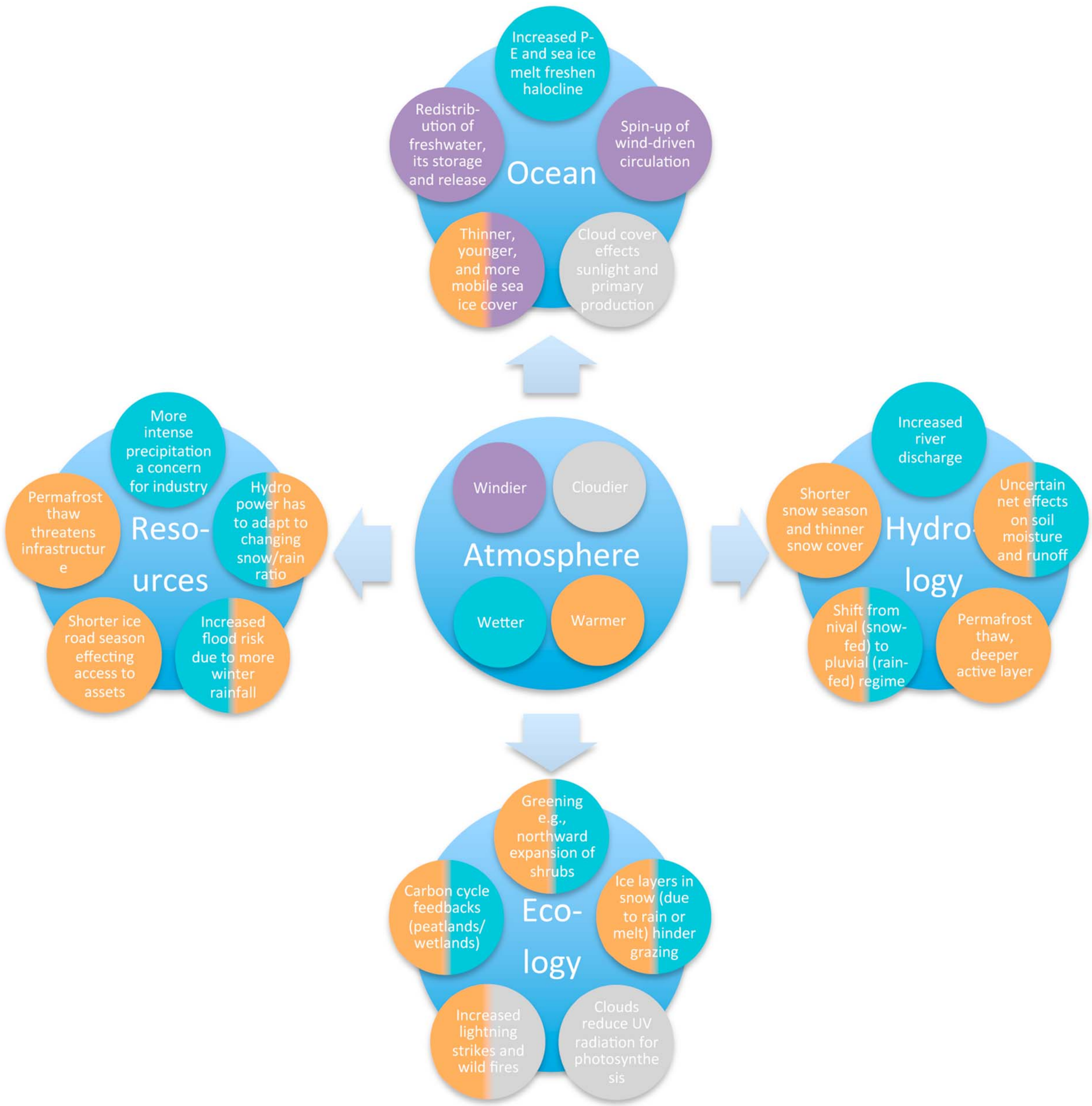

Figure 7. Schematic summarizing some of the important impacts of an intensified Arctic atmospheric water cycle. Impacts are colored depending on whether they are primarily a result of a warmer (orange), wetter (blue), windier (purple), and cloudier (grey) atmospheric state. This is not intended to be comprehensive, but rather to illustrate the multifaceted nature of change in the Arctic freshwater system.

et al., 2009; *Polyakov et al.*, 2012]. In coastal regions, the reduction of sea ice cover and related increase in wind forcing on the open ocean have increased wave activity and erosion rates of permafrost coasts [*Overeem et al.*, 2011].

Irrespective of changes in wind, the loss of sea ice (or more specifically, the increase in leads and open water) has allowed for greater wind forcing of the upper ocean and sea ice, with important implications on shelf-break upwelling [e.g., *Carmack and Chapman*, 2003]. The sea ice cover is becoming more mobile, mostly due to thinning and mechanical weakening of ice [*Rampal et al.*, 2009; *Vihma et al.*, 2012], although increase in wind forcing has played a role in the Central Arctic but not in the entire basin [*Spreen et al.*, 2011]. The more mobile sea ice cover affects the freshwater redistribution within the Arctic and export from the Arctic [*Carmack et al.*, 2016].

Changes in phase of precipitation have implications for the sea ice albedo and hence, mass balance. *Screen and Simmonds* [2012] estimated that summer snowfall has declined over the Arctic, lowering the ice albedo during the melt season (snow-covered ice is more reflective than bare ice). This albedo decline due to less snow on ice was estimated to be comparable to that related to the decline in sea ice cover [*Screen and Simmonds*, 2012]. Accordingly, the decrease in summer snowfall and increase in rain have probably had a large contribution to the Arctic sea ice decline. Models project future declines in snow on ice both in summer and spring, the latter due to later autumn freeze-up allowing less time for snow accumulation on ice [*Hezel et al*., 2012]. The declines reduce the surface albedo, and a lower albedo enhances surface melt leading to ponding, which further lowers the albedo. On the other hand, if winter snowfall increases, the snowpack gets thicker, which causes delay in the spring-summer decrease of albedo, reducing the accumulated melt.

The interactions between the oceanic and atmospheric components of the Arctic freshwater system are reciprocal. Changes in sea ice and sea surface temperature affect ocean-atmosphere moisture and energy fluxes, with local and remote atmospheric impacts [e.g., *Deser et al*., 2010; *Screen et al*., 2013]. Changes in river discharge and net precipitation alter salinity, and hence buoyancy, affecting ocean circulation and SSTs [*Carmack et al.*, 2016] and thereby, the atmosphere. An example is the projected cooling of the atmosphere over eastern Greenland and the northern North Atlantic during the 21st century due to increased freshwater in the North Atlantic and weakening of the thermohaline circulation [e.g., *Drijfhout et al*., 2012].

5.3. Hydrological Impacts

The impacts of changes in atmospheric water cycle on terrestrial hydrology include considerable spatial variability between and within river basins [*Bring et al*., 2016]. Many impacts are also sensitive to the exact time period studied. Here we only summarize the major large-scale impacts during the recent decades and some projections for the future. Increased precipitation over the Arctic river catchments (Figure 7) has been associated with greater discharge from the large Arctic rivers [*Peterson et al*., 2002; *Richter-Menge and Overland*, 2009]. This increase in river flow is mostly due to increased net precipitation.

Precipitation and its division between snow and rain affect the extent and thickness of terrestrial snow cover (Figure 7), which are important for ecology [*Wrona et al*., 2016]. According to *Brown and Derksen* [2013], who compared four independent sources of snow cover data, Eurasian autumnal snow cover extent has decreased over 1982–2011. Also, autumn precipitation has increased in East Siberia and the rain-to-precipitation ratio has increased in autumn in southern Canada and Eurasia south of Siberia (section 3.1.3). These changes affect autumn soil moisture and may have implications for spring runoff response and extreme flood events. In winter, atmospheric moisture, precipitation, and snow accumulation have increased over large parts of Russia [*Bulygina et al*., 2009; *Borzenkova and Shmakin*, 2012; *Callaghan et al*., 2011; *Park et al*., 2012]. Higher temperatures in spring and early summer have resulted in earlier melt, and the reduction of Northern Hemisphere snow cover in June has been even faster than the reduction of Arctic sea ice cover in September [*Derksen and Brown*, 2012].

Projected changes in precipitation phase can cause a shift from nival (snow-fed) to pluvial (rain-fed) regimes and with important consequences for the terrestrial freshwater system [see *Bring et al*., 2016] and ecosystems [*Wrona et al*., 2016]. Projected changes in precipitation and evapotranspiration also influence soil moisture and runoff; however, nonatmospheric factors also play a role. Warming contributes to permafrost thaw and a deepening of the active layer [*Bring et al*., 2016], with implications for infrastructure [*Instanes et al*., 2016].

The Greenland ice sheet is an indicator of long-term changes in the regional freshwater cycle. For estimates on the mass loss and runoff from Greenland and their effects on the sea level rise, see *Bring et al*. [2016]. Snowfall is the main source term for the mass balance of the ice sheet. Also, rain acts as a source term, as the rain water mostly freezes in the ice sheet, but the mass gain from rain is negligible compared to snowfall. The export of blowing snow and the sum of evaporation, sublimation, deposition, and condensation are small terms [*Box et al*., 2004; *Lenaerts et al*., 2012]. The main sinks of mass are the surface meltwater and glacier discharges, the latter including iceberg calving and glacier flow into the sea. Clouds are essential for the surface melt, as the cloud radiative forcing is positive (i.e., cloud have a net warming effect at the surface) over most of Greenland even in summer [e.g., *Bennartz et al*., 2013] due to the high surface albedo. This differs from conditions elsewhere in the same latitudes, where cloud radiative forcing is negative for several months during the warm season, and also from conditions over sea ice in the central Arctic (section 2.2).

During the past decade, the Greenland ice sheet mass balance has been negative due to increasing glacier discharge and meltwater runoff [*Box et al*., 2012; *Franco et al*., 2013]. Increasing air temperatures are responsible for the increasing trend in meltwater discharge. Increasing surface melt also affects glacier dynamics and subglacial sliding. In addition to the trend, Greenland melt has a substantial interannual variation. The variations are related to large-scale atmospheric circulation (characterized by NAO and the Greenland Blocking Index) [*Hanna et al*., 2014], associated transport of moisture over Greenland [*Neff et al*., 2014], and changes in cloud cover and properties [*Bennartz et al*., 2013]. Climate model experiments suggest that in a warmer climate there will be more winter snowfall, but the mass gain will not compensate the increasing mass loss via increased summer meltwater runoff [*Fettweis et al*., 2013]. Also, smaller Arctic glaciers and ice caps have continued to lose area and mass during the latest decades [*Vaughan et al*., 2013].

5.4. Ecological Impacts

Ecological impacts of changes in clouds, precipitation, and evaporation can be divided into impacts on marine [*Carmack et al*., 2016] and freshwater ecosystems [*Wrona et al*., 2016]. For the terrestrial freshwater ecology (Figure 7), for example, changes in temperature and precipitation may partially drive greening (e.g., poleward expansion of shrubs) or browning; and increased winter rainfall in a warmer climate (and associated ice layers in snow) affects the ecosystem [*Wrona et al*., 2016]. The fluxes of carbon dioxide and methane between the Earth surface and atmosphere are mostly controlled by processes at or below the Earth surface, such as changes in water and sediment temperatures, thawing of subsea and terrestrial permafrost, and changes in water budgets and levels [*Wrona et al*., 2016]. These processes are, however, strongly affected by accumulated changes in the air temperature and precipitation.

For both marine and freshwater ecosystems, potential changes in cloud cover are important as they affect the intensity of solar radiation, including UV radiation. The effects of UV radiation on Arctic freshwater ecosystems have been reviewed by *Wrona et al*. [2006]. The potential harmful effects include biomolecular, cellular, and physiological damage to organisms, and alterations in species composition. An increase in cloud-to-ground lightning strikes over the Arctic region is projected in a warming climate. Correspondingly, the burned area is expected to increase, especially over high latitudes [*Krause et al*., 2014].

5.5. Implications for Freshwater Resources and Human Activities

The atmospheric water cycle has a large direct (e.g., flooding) and indirect effect on human activities in the Arctic (Figure 7), as precipitation and evaporation affect the soil water budget and the thickness and extent of snowpack, and clouds affect the net radiation and, hence, the Earth surface temperature. The status and changes in soil water, snowpack, and surface temperature are essential for numerous economic activities, such as agriculture, forestry, hydropower, reindeer herding, road, and rail transport, and in particular in the permafrost zone, for construction work and infrastructure maintenance. Of particular challenge are the effects of extreme weather events. Recent changes in changes in the frequency and intensity of precipitation extremes have been spatially variable (section 3.1.3), but precipitation extremes are projected to increase over midlatitude and high latitude (section 4.1.3) and rain events in winter will become more common [*McAfee et al*., 2013], causing challenges, among others, to hydropower production [*Instanes et al*., 2016]. Hence, regional downscaling models have been developed at least for Norway, Iceland, Sweden, and Greenland to capture the extreme events [*Instanes et al*., 2016]. Another important aspect for hydropower production are the latitudinal differences in changes in net precipitation, the high latitudes becoming more water rich while the more southerly latitudes are becoming more “water poor” [*Prowse et al*., 2015b].

Precipitation over the sea also affects the snowpack on sea ice and, via snow-to-ice transformation, the thickness and structure of sea ice itself, with impacts on navigation. The effects of atmospheric water cycle on human activities via impacts on the Earth surface conditions are more discussed in *Bring et al*. [2016], *Instanes et al*. [2016], and *Wrona et al*. [2016].

More direct effects of atmospheric water cycle include effects of changes in cloud conditions on aviation (low clouds and fog are the main problems in the Arctic), atmospheric icing of structures (wind mills [*Makkonen et al.*, 2001], towers, and power lines), and on solar shortwave and thermal longwave radiation. A reduced cloud cover results in an increase in solar radiation, which has a strong positive impact on vitamin D status [*Andersen et al.*, 2013] and mental health of people [*Grimaldi et al.*, 2009] but simultaneously increases the occurrence of skin cancer [*De Fabo and Noonan*, 2002]. On the other hand, in the Arctic and sub-Arctic the cloud radiative forcing is positive for a large part of the year, so that an increase in cloudiness yields higher near-surface air temperatures, which reduce mortality [*Kue Young and Mäkinen*, 2010] and improve well-being and physical performance of people [*Rintamäki*, 2007].

6. Major Knowledge Gaps and Future Research Directions

Our knowledge of the Arctic freshwater cycle has gaps related to limited understanding of physical processes, insufficiency and inaccuracy of observations, and deficiencies in modeling. These generate uncertainty about the past and future changes in the water cycle.

6.1. Process Understanding

The Arctic freshwater cycle is affected by numerous small-scale processes in the atmosphere, snow, sea ice, and the ocean. Many of these processes and their parameterization in weather and climate models were recently reviewed by *Vihma et al*. [2014]. Hence, we only provide a very brief summary here. Insufficiently understood atmospheric processes include interactions of aerosols, cloud droplets, and ice particles in mixed-phase clouds as well as in the coupling of clouds, radiative transfer, and turbulence [*Shupe et al*., 2013; *Treffeisen et al*., 2007; *Morrison et al*., 2012]. In addition, there are major knowledge gaps in circumpolar-scale feedbacks in the Arctic, above all related to the interaction of water vapor and cloud feedbacks with the other feedbacks contributing to the Arctic amplification of climate warming [*Serreze and Barry*, 2011; *Pithan and Mauritsen*, 2014]. Lack of process understanding is due to complexity of the processes and limited amount and accuracy of observations available.

6.2. Insufficient Observations

The main reason for insufficient knowledge on the past changes and present status of the atmospheric water cycle in the Arctic is the lack of spatially well-covered and accurate data. This is most serious over the oceans, but problems also remain over land areas, with a coarse station network. Considering different atmospheric water variables, the lack of in situ observations is most evident for evaporation and precipitation [*Yang et al*., 2005; *Pavelsky and Smith*, 2006]. Further, there are very few data on cloud properties (such as cloud water and ice contents and droplet size distribution) and air moisture above the atmospheric surface layer. Except for limited short-term campaigns, there are no in situ observations of the above mentioned variables over the Arctic and sub-Arctic seas. Over land areas, vertical profiles of air moisture are routinely measured at only some 36 radiosonde sounding stations north of 65°N [*Nygård et al*., 2014]. Important advances have been made in ground-based remote sensing of air moisture and cloud properties, but measurements are only made at a few stations in the circumpolar Arctic and the records are still short [*Shupe*, 2011; *Shupe et al*., 2011]. During the last decades, satellite remote sensing has started to provide important information on air moisture and clouds, in particular from microwave sensors [*Bobylev et al*., 2010], but optical data on clouds are missing from the Arctic night.

A large part of the existing data is very uncertain. In situ observations of snowfall may include errors of up to 200%, mostly due to the effect of wind [*Aleksandrov et al*., 2005], and new satellite-borne sensors of precipitation (e.g., the Global Precipitation Monitoring satellite) provide coverage only up to about 65°N. Observations and estimates of evapotranspiration over terrestrial surfaces suffer from various errors and uncertainties [*Bring et al*., 2016]. Accuracy of observations of air humidity is poor in low temperatures; unheated and unventilated hygrometers often become coated with ice [*Déry and Stieglitz*, 2002; *Makkonen and Laakso*, 2005] and the accuracy of radiosonde data decreases as the water vapor concentration and pressure decrease [*Elliot and Gaffen*, 1991]. Cloud observations are mostly visual and liable to large errors during the Arctic night [*Hahn et al*., 1995]. The accuracy of satellite remote sensing observations on cloud water and ice content is reduced by problems in distinguishing the signals originating from the snow/ice-covered Earth surface and from the atmosphere, in particular in the case of low clouds. The advent of active satellite sensors (e.g., the A-Train)

[*L'Ecuyer and Jiang*, 2010] improved this situation, but the time series are very short and only cover the southern Arctic (< ~82°N). The remote sensing problems are smaller for air humidity than for cloud condensate content, but the data suffer from a limited vertical resolution [*Boisvert et al*., 2015].

6.3. Deficiencies in Models and Reanalyses

Our knowledge on air moisture, clouds, precipitation, and evaporation in the Arctic is also limited by errors and inaccuracies in modeling of these variables, due to imperfect representation of complex physical processes (e.g., cloud microphysics) and coarse model resolution. The deficiencies are discussed more thoroughly in *Lique et al.* [2016], and here we focus on some of those related to the results presented in sections 3 and 4.

Considering numerical weather prediction (NWP) results and atmospheric reanalyses, the problems in modeling and observations are tied via data assimilation. The observational input to model analyses and reanalyses varies in time and has limited spatial coverage. Comparisons against in situ observations have demonstrated that atmospheric reanalyses have large errors in the vertical profiles of air temperature as well as specific and relative humidity over the Arctic Ocean, with predominantly warm and moist biases in the lowermost 0.5–1 km [*Lüpkes et al*., 2010; *Jakobson et al*., 2012; *Wesslén et al*., 2014]. The observed biases in humidity are in many cases comparable to or even larger than the climatological trends during the latest decades. Reanalyses also have problems in reproducing the observed cloud characteristics and cloud radiative forcing in the Arctic [*Walsh and Chapman*, 1998; *Wesslén et al*., 2014]. Further, reanalyses are not in hydrological balance, i.e., the annual mean vertically integrated water vapor flux convergence does not equal net precipitation in the Arctic. For the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) and NCEP/DOE reanalyses the imbalance was as large as 60% [*Cullather et al*., 2000]. There is also a considerable inaccuracy in the vertical profile of water vapor transport to the Arctic; radiosonde sounding data suggest that the transport peaks at 850 hPa level [*Overland and Turet*, 1994], but ERA-40 suggests the peak much closer to the Earth surface [*Jakobson and Vihma*, 2010]. We do not know which is closer to the truth; reanalyses are inaccurate but radiosonde data are biased to land areas.

Reanalyses as well as NWP and climate models also suffer from problems in understanding and parameterization of subgrid-scale physical processes related to moisture variables. Although there have been recent advances in understanding the physics of Arctic clouds [*Morrison et al*., 2012], these have not yet yielded remarkable improvements in model parameterizations [*Vihma et al*., 2014] and performance (section 3.2.1). Models still rely on parameterizations based on more well understood, lower latitude clouds. This may have contributed to the recent finding that radiation and air temperature errors of climate models in the Arctic are strongly related to errors in cloud occurrence and heights, as well as to the vertical distribution of cloud water and ice content [*Tjernström et al*., 2008; *Kay et al*., 2011; *Cesana et al*., 2012; *Liu et al*., 2012; *Wesslén et al*., 2014; *de Boer et al*., 2014]. Further, CMIP3 models suffer from a large intermodel spread, especially in winter, and some of the models even have an annual cycle with fewer clouds in summer than in winter, i.e., opposite to the observations [*Karlsson and Svensson*, 2011].

Considering Earth surface processes, sea ice [*Cheng et al*., 2008, 2013] and terrestrial hydrology models [*Lique et al.*, 2016; *Bring et al*., 2016] suffer from considerable inaccuracy in precipitation forcing. In particular, there is a strong need for improvement in modeling the distribution of precipitation between rain and snow. Further, stand-alone ocean models sometimes simply mimic the surface freshwater budget by relaxing the upper ocean salinity to climatology [e.g., *Stössel et al*., 2011]. There is also need for better parameterizations for evaporation from spray droplets [e.g., *Andreas et al*., 2008] and drifting/blowing snow [e.g., *Gordon et al*., 2006].

Due to the above mentioned problems, the past climatology on all moisture variables in the Arctic is highly uncertain, particularly over the ocean prior to 1979, when the assimilation of satellite data into reanalyses was started. Evidence for trends in Arctic precipitation is complicated by inadequacies and inaccuracies both in in situ and remote sensing measurements [*Walsh et al*., 2011b], and in the case of evaporation the situation is even worse. A common assumption is that evaporation has increased in a warmer Arctic, but also regional and seasonal decreasing trends have been reported [*Screen and Simmonds*, 2010b; *Boisvert et al*., 2013]. See *Carmack et al.* [2016] for more discussion on evaporation over Arctic Seas.

6.4. Uncertainty in Future Projections

Uncertainties in model projections arise from three sources: scenario uncertainty, model uncertainty, and internal variability [*Hawkins and Sutton*, 2009; *Lique et al*., 2016]. For near-term trends (out to approximately 2050), internal variability is a major source of uncertainty for regional climate projections [*Deser et al*., 2012, 2014]. Across-model differences are the other major source of uncertainty prior to about 2050 [*Hodson et al*., 2013]. This is especially the case for the Arctic region because of large natural variability of high-latitude climate [*Screen et al*., 2014]. Such large dependence of future trends on internal variability (for example, Arctic sea ice extent) [*Wettstein and Deser*, 2014] means that large ensembles of climate simulations, with each ensemble starting from a different initial condition, are required to accurately constrain forced trends. Model archives such as the CMIP5 often include only one, or a small number of realizations per model, and thus, it is difficult, if not impossible, to quantify the extent to which spread in near-term projections is due to internal variability or model uncertainty. Efforts to produce large ensembles of a single model are an important step forward in this regard [e.g., *Deser et al*., 2012, 2014; *Kay et al*., 2014] and have revealed the importance of accounting for internal variability. They have demonstrated that a large component of the CMIP5 model spread can be explained by internal variability. More such large ensembles with a greater diversity of models are required to better understand the sources of uncertainty in projections of the atmospheric water cycle.

On multidecadal to centennial time scales, internal variability is less important and model uncertainty dominates [*Hawkins and Sutton*, 2011]. For variables such as precipitation and evaporation, the latest models generally agree on the sign of the response to increasing GHG concentrations. However, there is a large model divergence of magnitudes of the projected changes and regional differences in sign and magnitude. For variables such as cloud cover and cloud properties, the models often disagree on the sign, seasonality, and spatial pattern of the response. This is not surprising, as the cloud response to sea ice loss is a sensitive balance. Sea ice loss results in increases in the surface fluxes of both sensible and latent heat, the former restraining and the latter favoring cloud formation. As a rule of thumb, there is greater confidence in projected changes that are closely tied to thermodynamics (e.g., warming) and less confidence in those related to dynamical changes (e.g., changes in atmospheric circulation, turbulence, and mixing). The latter is a topic ripe for future research [*Shepherd*, 2014].

6.5. Future Research Directions

To better quantify the Arctic atmospheric water cycle and its changes, there is a need for more extensive and accurate observations, better process understanding, better models, and more extensive and systematic utilization of existing models. Considering observations, recommendations for future research are naturally closely tied to logistical and economic resources available. The possibilities for a large increase in the number of radiosonde sounding stations in the Arctic are not high, but there may be potential for sustainable cost-effective vertical profiling of the lower troposphere applying small Unmanned Aerial Vehicles (UAVs) [e.g., *Jonassen et al*., 2012] and for at least occasional missions applying long-range UAVs and dropsondes [*Intrieri et al*., 2014]. In addition, ground- or ship-based remote sensing of cloud properties and air humidity has potential to yield significant further advances in process understanding [e.g., *Tjernström et al*., 2014] and, when the time series will become long enough, cloud climatology [*Shupe et al*., 2011; *Uttal et al*., 2015]. Further research should also address the observational and parameterization challenges of evaporation from drifting/blowing snow and spray droplets.

Analogous to the situation in which sea ice loss can produce local and remote effects on weather and climate [e.g., *Vihma*, 2014], there is the potential for similar effects to be produced by the significant reductions that have occurred, and are projected to continue, in lake and river ice coverage, as reviewed by *Bring et al*. [2016]. Of particular interest is the effect of ice loss and associated heating of freshwater bodies on the local evaporative flux and overall contribution to atmospheric moisture transport to and from the Arctic Freshwater System. The magnitude of such could be significant given the large spatial extent of ponds and lakes found at the midlatitude to high latitude [*Prowse et al*., 2011] areas that are comparable to those of sea ice loss that have occurred over recent decades [e.g., *Swart et al*., 2009].

Modeling issues that deserve more attention also include the representation of unresolved orographic effects on precipitation [*Cullather et al*., 2014], storm tracks, and atmospheric rivers [*Newman et al*., 2012], as well as large-scale effects of sea ice on evaporation and precipitation [*Zhang and Walsh*, 2006], and especially on

$P-E$ which is the key to land surface drying/wetting and impacts ocean salinity through runoff and directly. Model results for the recent climate and changes have shown various discrepancies, but identification of their exact causes is still difficult. Systematic evaluations on different aspects of the model simulations, such as in *Kattsov et al*. [2007], may yield more information. Considering future projections of the atmospheric water cycle, there is need for more model experiments with large ensembles and a great diversity of models. Systematic comparisons of the results should allow a better understanding of the sources of uncertainty. Further, there is need for updated information on the contributions of anthropogenic and natural forcing to changes in the Arctic freshwater cycle (Figure 4).

7. Conclusions

The physical processes related to atmospheric moisture, clouds, precipitation, and evaporation in the Arctic are complex. Although our understanding of individual processes has increased, challenges remain in the understanding of nonlinear interactions of various processes, partly acting at different spatial and temporal scales. This challenge in understanding is reflected in problems of modeling the processes, in particular the subgrid-scale processes that need to be parameterized in climate and numerical weather prediction models. Concrete advances in parameterizations have not been fast.

Past trends in atmospheric water variables in the Arctic include large spatial variations and are sensitive to the time period and source of data. This is the case particularly for precipitation, snowfall, evaporation/evapotranspiration, and cloud variables. Over longer periods and large areas, there seems, however, to be a trend toward wetter conditions, but when smaller regions and shorter time periods are analyzed, different trends are detected. Models capture the overall wetting trend but have problems in reproducing the regional details. The latter may partly result from the larger influence of interannual variability at regional scales and partly (e.g., in regions of complex orography) from the insufficient spatial resolution. The overall wetting is linked to the general warming trend, partly driven by anthropogenic forcing, and amplified in the Arctic due to several feedback effects. The wetter conditions reflect an intensification of the atmospheric water cycle in the Arctic, seen as increases in the moisture transport from lower latitudes and precipitation in the pan-Arctic scale, with a considerable uncertainty in the changes of evaporation in the Arctic.

Over approximately the next 100 years, climate model simulations indicate robust increases in precipitation and specific humidity. In the case of evaporation, the projected trends vary between regions and seasons, and for clouds the projections based on different climate models are diverse. A robust increase in precipitation with scattered results for trends in cloudiness seems possible if the clouds contain more water and the precipitation intensity accordingly increases. Projections for time scales of the order of years to a few decades have considerable uncertainty due to the large natural interannual and decadal variability. New and sustained observations and process-level observations, especially in data sparse regions such as the Arctic Ocean, are needed to better understand ongoing changes and to better evaluate and improve reanalyses and climate models. To narrow uncertainty in future projections, climate models need to be improved in close collaboration with observational community and more model experiments with large multimodel ensembles are needed.

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