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Key Points:

- We review processes across hydrophysiographic regions of the terrestrial Arctic freshwater system
- Arctic hydrologic change affects atmosphere, ecology, resources, and oceans
- Interfaces between hydrology and other Earth system components are critical

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Arctic terrestrial hydrology: A synthesis of processes, regional effects, and research challenges

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Abstract Terrestrial hydrology is central to the Arctic system and its freshwater circulation. Water transport and water constituents vary, however, across a very diverse geography. In this paper, which is a component of the Arctic Freshwater Synthesis, we review the central freshwater processes in the terrestrial Arctic drainage and how they function and change across seven hydrophysiographical regions (Arctic tundra, boreal plains, shield, mountains, grasslands, glaciers/ice caps, and wetlands). We also highlight links between terrestrial hydrology and other components of the Arctic freshwater system. In terms of key processes, snow cover extent and duration is generally decreasing on a pan-Arctic scale, but snow depth is likely to increase in the Arctic tundra. Evapotranspiration will likely increase overall, but as it is coupled to shifts in landscape characteristics, regional changes are uncertain and may vary over time. Streamflow will generally increase with increasing precipitation, but high and low flows may decrease in some regions. Continued permafrost thaw will trigger hydrological change in multiple ways, particularly through increasing connectivity between groundwater and surface water and changing water storage in lakes and soils, which will influence exchange of moisture with the atmosphere. Other effects of hydrological change include increased risks to infrastructure and water resource planning, ecosystem shifts, and growing flows of water, nutrients, sediment, and carbon to the ocean. Coordinated efforts in monitoring, modeling, and processing studies at various scales are required to improve the understanding of change, in particular at the interfaces between hydrology, atmosphere, ecology, resources, and oceans.

1. Introduction

Many of the environmental changes currently under way in the Arctic involve terrestrial freshwater. Central examples include a decreasing extent and duration of snow cover [*Callaghan et al.*, 2011; *Vaughan et al.*, 2013], increasing flow from large Siberian rivers [*Peterson et al.*, 2002, 2006; *McClelland et al.*, 2006; *Shiklomanov and Lammers*, 2009; *Overeem and Syvitski*, 2010] and glaciers and ice sheets [*Mernild and Liston*, 2012; *Church et al.*, 2013], and changing partitioning between surface water and groundwater [*L. C. Smith et al.*, 2007; *Walvoord and Striegl*, 2007]. These changes illustrate that the Arctic Freshwater System (AFS) also comprises feedbacks between the terrestrial hydrology and the ocean, atmosphere, ecosystems, and natural resources [*Hinzman et al.*, 2013; *Raymond et al.*, 2013].

Over the past few decades, investigation efforts into the AFS have increased rapidly, and a large body of research, as well as a number of synthesis reports, has made major contributions to our knowledge. Nevertheless, our understanding of many of the changes remain incomplete, particularly with respect to the effects on and feedbacks to the ocean, atmosphere, ecosystems, and natural resources and also with respect to the dynamics with which the ongoing rapid changes will unfold across various regions of the pan-Arctic hydrological drainage basin.

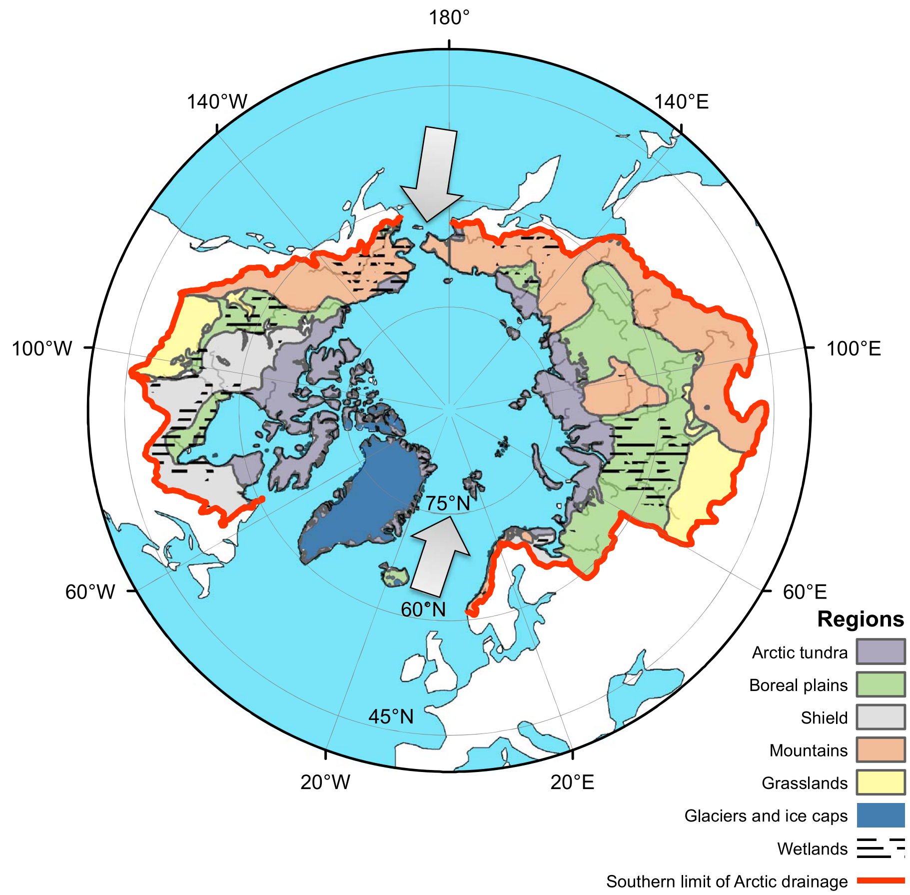

Figure 1. Overview of regions. Schematic map (polar stereographic projection) illustrating a number of major hydrophysiographic regions in the terrestrial pan-Arctic drainage basin. The outer boundary of the basin is drawn to consider all areas potentially contributing to Arctic Ocean freshwater inflow, when also including ocean freshwater transport through the Bering Strait and the North Atlantic (indicated with arrows; see discussion on contributing area in *Prowse et al.* [2015a]. Although we emphasize that there is substantial variation within the regional boundaries outlined here, and many details that are excluded in this map, these large regions highlight a set of overarching landscape types across the Arctic drainage basin.

This paper is a part of the Arctic Freshwater Synthesis (AFS), a review effort initiated by the Climate and Cryosphere (CliC) project of the World Climate Research Program (WCRP), the International Arctic Science Committee (IASC), and the Arctic Monitoring and Assessment Program (AMAP), to provide updated information about the AFS. The AFS introduction paper [*Prowse et al.*, 2015a] contains background and can be consulted for a detailed discussion of the study domain, the pan-Arctic drainage basin. The AFS also comprises four other component papers that focus on freshwater processes related to the ocean [*Carmack et al.*, 2016], atmosphere [*Vihma et al.*, 2016], ecosystems [*Wrona et al.*, 2016], and natural resources [*Instanes et al.*, 2016], respectively. A fifth AFS component paper addresses modeling issues related to the AFS [*Lique et al.*, 2016], and a summary paper [*Prowse et al.*, 2015b AFS] concludes the synthesis with key points and recommendations from the entire AFS.

In this AFS component paper, we provide a synthesis of terrestrial Arctic hydrology processes, their change drivers, and the main research challenges associated with them. Furthermore, a main aim is to identify linkages between terrestrial hydrology processes and other components of the AFS, and we therefore refer to other AFS papers throughout. We also attempt to separate distinctive aspects of freshwater processes, fluxes, and storages, and their changes, across representative hydrophysiographic regions of the AFS study domain, the pan-Arctic drainage basin (Figure 1), something that has received less focus in previous investigations (for brevity, the term Arctic is used in this paper to refer to the AFS domain).

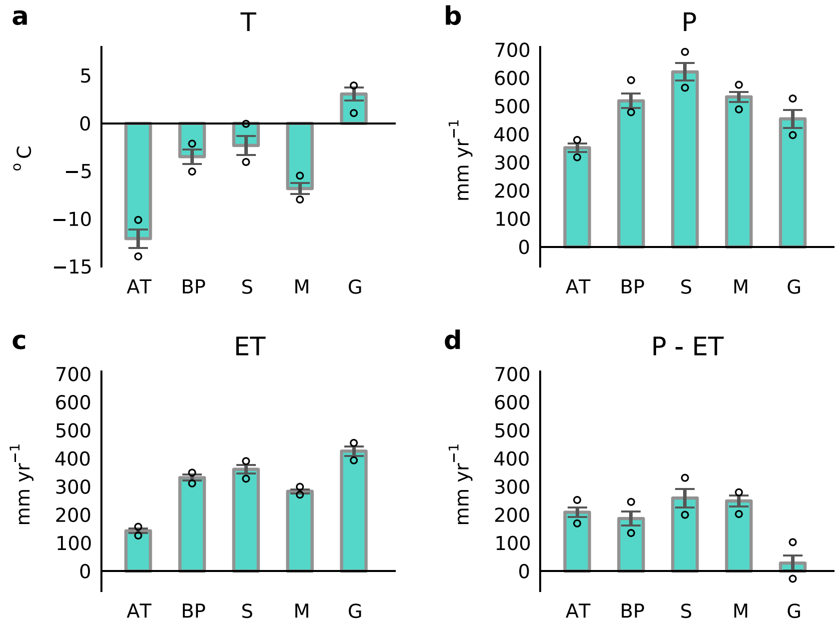

Figure 2. Climatology of Arctic hydrophysiographical regions. Annual averages of grid-based (a) temperature, (b) precipitation, (c) evapotranspiration, and, as a proxy for runoff, (d) precipitation minus evapotranspiration values, over the regions defined in Figure 1 for the period 1979–2013. Regions are abbreviated as AT: Arctic tundra, BP: Boreal plains, S: Shields, M: Mountains, and G: Grasslands. Error bars denote one standard deviation of annual means, and open circles denote maximum and minimum annual means during the period. All data are from the ERA-Interim reanalysis product and are available at http://apps.ecmwf.int/datasets/data/interim-full-daily. Glaciers and ice sheets are not included here due to low density of observations in these ecoregions.

The hydrophysiographic regions we use in this paper mostly correspond to the terrestrial ecoregions defined by the World Wide Fund for Nature [*Olson et al*., 2001]. The ecoregions (Arctic tundra, boreal plains, grasslands, and glaciers/ice caps) are used throughout the AFS, and are discussed more in detail in *Prowse et al*. [2015a]. However, due to the distinct hydrological and hydrogeological properties of mountains, shield regions, and wetlands, we also include those three categories as additional hydrophysiographic regions in this AFS component. The data sets used to define these regions are, for mountains, a study by *Adam et al*. [2006], for shield regions, the Canadian Geological Survey [*Kirkham et al*., 1995], and for wetlands, the Global Lakes and Wetlands Database [*Lehner and Döll*, 2004]. We stress that each of these regions extends over a large area and therefore comprises quite diverse environments where local variations in hydrology are likely to be substantial. However, despite inevitable ambiguity and internal diversity, each region has still been identified as distinct in the aforementioned sources, and we use them here as overarching categories of hydrophysiographical landscape types.

2. System Functioning and Key Processes

This section presents a short description of the main processes pertaining to the Arctic terrestrial freshwater system, with focus on freshwater storages and fluxes. We discuss precipitation, evapotranspiration, surface runoff and channel flows, permafrost and groundwater hydrology, and river and lake ice. Following the processes, we highlight the central characteristics of a number of Arctic hydrophysiographical regions, also with regard to freshwater, its main storages, and fluxes. These regions are outlined in Figure 1, and an overview of regionally averaged temperatures and water fluxes based on gridded data is presented in Figure 2 (see also further discussions in *Lique et al*. [2016], *Prowse et al*. [2015a], and *Vihma et al*. [2016].

Throughout the remainder of the paper, we progress from describing the system (section 2) to treating its past and future changes (sections 3–4). We then review key linkages across components of the Arctic freshwater system (section 5) and finally highlight research needs (section 6). As in this section, all of the following sections 3–6 are separated into a first section where we discuss processes (with some regional examples occasionally highlighted) and a second section where we review particular effects in each hydrophysiographical region. With this organization, we hope that a reader interested in specific processes or regions is able to quickly find the right material in the paper.

2.1. Processes

Precipitation is the major flux into the terrestrial freshwater system (see *Lique et al*. [2016], *Prowse et al*. [2015a], and *Vihma et al*. [2016] for overviews and quantitative estimates). A substantial proportion of Arctic annual precipitation is falling and stored as snow, of heterogeneous spatial distribution [e.g., *Liston and Hiemstra*, 2011; *Mernild et al*., 2014], and released to the river network in a relatively short window of time during spring snowmelt. Regionally, precipitation stored as snow contributes to runoff also in summer months. The phase of precipitation influences both annual and seasonal water balances at multiple spatial scales. The intensity with which precipitation is delivered also impacts runoff generation, with the spring freshet occasionally rivaled by summer thunderstorms for smaller basins [*Kane et al*., 2008].

Over most Arctic basins, the majority of precipitation returns to the atmosphere as evapotranspiration (ET). ET links the water and energy cycles and couples the land to the atmosphere (evaporation over the Arctic Ocean is treated in more detail in *Vihma et al*. [2016]). The Arctic annual ET water flux is generally smaller than annual precipitation, except over a few southern inland areas [*Serreze et al*., 2006, Figure 1] and over lakes and wetlands, where summer ET may exceed summer, or even annual, precipitation [*Marsh and Bigras*, 1988; *Rovansek et al*., 1996; *Bowling and Lettenmaier*, 2010]. As the transpiration depends on the vegetation canopy, ET varies considerably even on local scales, as well as in time. The length of the snow cover and growing seasons are critical controls on the ET water flux. In addition, landscape variations such as lake area change may add up to considerable effects on ET and runoff [*Hinzman et al*., 2005; *Karlsson et al*., 2015]. Large-scale evapotranspiration values are difficult to estimate, but recent satellite-based assessments indicate pan-Arctic averages of ~230 mm yr^{-1} [*Zhang et al*., 2009], with values ranging from 136 mm yr^{-1} in grasslands to 596 mm yr^{-1} in evergreen broadleaf forests [*Mu et al*., 2009]. The ET data collected locally at eddy flux towers have been widely used for parameterization of hydrologic models [*Sun et al*., 2008] and ET algorithms for remote sensing products [*Ruhoff et al*., 2013]. Airborne eddy correlation measurements provide a useful means of bridging the scale interval between flux towers and regional models [*Sellers et al*., 2012].

Besides ET, river discharge is the other major water flux out of Arctic basins. Freshwater flow through the Arctic's principal rivers conveys water, heat, sediments, carbon, and nutrients to the coastal domain and to the Arctic Ocean. Furthermore, along the river pathways, river flows, ice conditions, and runoff regimes control winter transportation, infrastructure and resource extraction, and ecosystem dynamics. Most of the water is transported to the Arctic Ocean during the spring snowmelt and in summer. The peak flow rates (May–June) can exceed the mean annual flow rate as much as 40 times for the Yenisey and Lena Rivers; the corresponding ratio for the Mackenzie River is much less, about 5 times [*Aagaard and Carmack*, 1989; *Bowling et al*., 2000]. Due to dam construction, shifts of discharge strongly influence the seasonal flow for many Arctic rivers, including the Yenisey, Lena, and Mackenzie [*McClelland et al*., 2004; *Yang et al*., 2014]. In general, the variability is dampened by regulation, as spring and summer flow is held back and released in winter when flows are low. Effects on annual flows, however, are limited for most of the rivers [*McClelland et al*., 2004; *Yang et al*., 2004b; *Stuefer et al*., 2011]. The largest Arctic rivers maintain channel flow year around including the winter flows under seasonal ice cover. Smaller northern rivers, however, often freeze to the bottom with no channel flow in winter. A portion of winter discharge is seasonally stored as river and lake ice and released in spring [*Prowse et al*., 2011]. Formation and growth of aufeis during winter are typical for many northern rivers [*Kane*, 1981]. In total, rivers deliver around 4×10^3 km^3 of freshwater annually to the Arctic Ocean [*Serreze et al*., 2006; *Haine et al*., 2015], although this figure is strongly dependent on the total contributing area that is considered [*Prowse et al*., 2015a].

Apart from the water itself, river conveyance of heat influences inland ice and ecosystem dynamics, particularly during transition seasons. The total delivery of water constituents such as nutrients, sediment, and carbon is less well known than that of the freshwater itself [*Bring and Destouni*, 2009], but recent estimates indicate fluxes of total nitrogen amounting to 1.3 Tg N yr^{-1} [*Holmes et al*., 2011], total phosphorus 73 Gg P yr^{-1} [*Holmes et al*., 2011], sediment 324–884 Tg yr^{-1} [*Hasholt et al*., 2006], inorganic carbon 57 ± 9.9 Tg C yr^{-1} [*Tank et al*., 2012], and dissolved organic carbon 34–38 Tg C yr^{-1} [*Holmes et al*., 2011]. The water chemical composition is often strongly controlled by landscape and ecosystem processes [*Koch et al*., 2013; *Aiken et al*., 2014] (see *Wrona et al*. [2016], for further discussion). Carbon from rivers is a key input to near-coastal and ocean acidification, processes which are further discussed in *Carmack et al*. [2016].

Water chemistry and water flows in Arctic basins are both often influenced by permafrost. Active layer dynamics govern a wide range of surface and subsurface processes across permafrost landscapes and control mechanisms of runoff generation. The topographic relief, which varies between mountains, slopes, and flat terrain, is also strongly influencing soil moisture. *Carey and Woo* [2001], *Quinton et al*. [2005], and *Semenova et al*. [2013] have shown the impact of topography, soil and vegetation on ground freeze-thaw, soil moisture storage, surface and subsurface flow distribution, evaporation, and other processes. Due to the large extent of the area underlain by permafrost, the active layer thickness (ALT) and behavior varies across the Arctic, which influences soil moisture and storage. The mechanical stability of soil is also influenced by the water and ice content of the active layer [see *Instanes et al*., 2016].

On shorter time scales, seasonal changes in surface ice are prominent characteristics of Arctic river systems. During winter, lake and river ice grow to cover $1.7 \times 10^6\,km^2$, an area approximately equal to the Greenland ice sheet. The peak volume of $1.6 \times 10^3\,km^3$ roughly matches the Northern Hemisphere snowpack on land [*Brooks et al*., 2013]. This freshwater ice produces numerous effects on physical systems, ecosystem services, and socioeconomic systems within Arctic freshwater storage and flow networks [*Instanes et al*., 2016; *Wrona et al*., 2016], although the hydrologic controls of many effects originate well outside sub-Arctic latitudes, via the headwaters of the large northward flowing rivers [*Bennett and Prowse*, 2010].

Most meteorological and climatological effects of variations in freshwater ice (e.g., coverage and duration) are confined primarily to the local or regional scale (e.g., radiation and convective fluxes), with the greatest effects produced by ice cover on large lakes [*Rouse et al*., 2005], although the process of river ice breakup has also been shown to be important, especially on the spring climate of large river deltas [*Prowse et al*., 2011]. However, effects on evapotranspiration, precipitation, and their feedbacks on hydrology may act on much larger scales, even influencing water balance of large Arctic basins [*Rouse et al*., 2008]. The magnitude and timing of hydrologic extremes such as low flows and floods are mostly controlled by the dynamics of river ice freezeup and breakup [*Beltaos and Prowse*, 2009], although reservoir discharge and groundwater base flow also control winter low flows [*Woo and Thorne*, 2014]. Spring breakup tends to be the dominant hydrologic event across the full domain of large rivers and deltas, such as the Mackenzie and the Lena [*de Rham et al*., 2008; *Goulding et al*., 2009; *Fedorova et al*., 2015], and also the main agent of sediment transport and morphological change [*Turcotte et al*., 2011].

2.2. Hydrophysiographic Regions

Here we summarize hydrological characteristics across the seven hydrophysiographic regions shown in Figure 1: Arctic tundra, boreal plains, shields, mountains, grasslands, glaciers and ice sheets, and wetlands. We also highlight a number of ways in which the hydrological responses to environmental changes can be expected to differ across the regions.

In the tundra, continuous permafrost, with an active layer that decreases from about 1.5 m in the south to 0.5 m or less in the north (Circumpolar Active Layer Monitoring data, available at http://www.gwu.edu/~calm/data/north.html), strongly influences water fluxes and storage. With limited permeability of the frozen soil (both seasonal and perennial) that restricts infiltration and water storage, groundwater storage and circulation are largely confined to the seasonally thawed zone. Generally, watersheds with a high percentage of permafrost coverage have low subsurface storage capacity for liquid water and thus a low winter base flow (minimum flow) and a high spring or summer peak flow (maximum flow) [*Woo*, 1986; *Kane*, 1997; *Yang et al*., 2002; *L. C. Smith et al*., 2007; *Ye et al*., 2009]. Evapotranspiration and precipitation are both very low, although evapotranspiration can be highly variable in shrublands [*Mu et al*., 2009]. At its northernmost extreme, the tundra is adjacent to the ocean, where groundwater and smaller streams transport relatively poorly quantified amounts of water, nutrients, sediment, and carbon to the ocean.

In boreal plains, the low gradients lead to slow surface and subsurface water movement and ample depression storage, which is associated with the extensive occurrence of wetlands [*Ferone and Devito*, 2004]. Permafrost becomes discontinuous in the south, which allows groundwater to circulate more freely in relation to the impervious frozen substrate in the southern boreal plains [*K. B. Smith et al*., 2007]. With increasing solar radiation and vegetation cover, evapotranspiration flux increases southward [*Mu et al*., 2009]. Seasonal storage of snow is less pronounced than in the tundra, and high flow events arising from rain become more pronounced southward, next to the spring freshet [*Su et al*., 2005].

In shield regions, the bedrock matrix is essentially impermeable but may be riddled with joints and fractures that, depending on their content, permit infiltration [*Spence and Woo*, 2002]. Thus, the runoff ratio can vary greatly on small scales. The major structural fissures are eroded into soil-filled valleys with lakes and wetlands that form tortuous drainage networks [*Spence and Woo*, 2003]. Groundwater yield is low relative to surface water, as permafrost and bedrock allow limited groundwater storage. Instead, surface storage in lakes and wetlands is central to runoff generation, which follows the principle of fill and spill, whereby lakes and wetlands in valleys have to be filled above the elevation of their outlet levels before flow commences [*Woo and Mielko*, 2007]. Otherwise, water is held back in storage and discharge is interrupted, leading to cessation of flow downstream.

Mountainous regions are responsible for >60% of the annual flow of Mackenzie Basin [*Woo and Thorne*, 2003] and are important freshwater sources also for larger basins in eastern Siberia and the Russian Far East. The Lena River receives on average 40% of the annual flow from the mountainous regions of the Aldan and Upper Lena Rivers [*Berezovskaya et al.*, 2005]. The large altitudinal range of mountainous regions leads to a prominent vertical zonation and an aspect-controlled microclimate. In high elevations, and at high latitudes also at lower elevation, water is stored interannually in snowpack and glaciers, with gradual melt release prolonging the duration of flow generation over the snow-free season. Orographic effects generally yield high precipitation, particularly in coastal regions [*Mernild et al.*, 2015]. Aspect plays an important role in the energy and water balances of slopes and tributary basins. Snowmelt on south facing slopes can be a month in advance of north facing slopes, and evapotranspiration is higher on south than north slopes. For example, for Wolf Creek, Yukon, evapotranspiration on a north slope averaged 315 mm yr^{-1} and on the opposite slope it reached 372 mm yr^{-1} [*Carey and Woo*, 2001].

Some large Arctic rivers pass through prairies (e.g., the Saskatchewan River that joins the Nelson River) and steppes (e.g., the Ob) in their upper courses. These low-relief areas allow air masses and their accompanying disturbances to sweep uninterrupted over great distances. This leads to large climatic variability that impacts terrestrial hydrology, principally with regard to floods and droughts. In addition to flooding from convectional rainfall, mountain rivers also effectively deliver snowmelt floods to the plains. In summer, low precipitation and high temperatures combine to effect large net water loss to evaporation. However, evaporation can be very variable due to large fluctuations in climate and surface conditions [*Armstrong et al.*, 2015]. Within this region, some basins have no outlet and form hydrological enclaves, disconnected from the surrounding watershed. During dry years, the low precipitation and decrease in snowpack lead to drying out of local wetlands [*Fang and Pomeroy*, 2008], which then become disconnected from each other [*Shaw et al.*, 2012].

Glaciers and ice sheets play a major role in the Arctic freshwater system through their impact on the surface energy budget, the water cycle, and sea level [*Vaughan et al.*, 2013]. Freshwater discharge from the Greenland ice sheet and the Arctic terrestrial rivers, the latter substantially larger, has a critical influence on Arctic Ocean circulation [see *Carmack et al.*, 2016; *Vihma et al.*, 2016; *Lique et al.*, 2016]. For smaller Arctic glaciers and ice caps, their role in freshwater storage and flux varies greatly throughout the region. With the exception of the Yukon River basin, glaciers contribute only a small share of flow for all the major Arctic rivers [*Dyurgerov et al.*, 2010]. Regionally, however, they strongly affect both runoff seasonality and water storage change, as freshwater flow is redistributed to the summer.

Wetlands are prominent through the Arctic, especially in the tundra, where they can function as large systems of lakes and wetlands, with water accumulating in depressions. There are different kinds of wetlands in the Arctic, with a central distinction differentiating between peatlands and wetlands in the general sense. In boreal western Siberia and Canada, peatlands are more extensive, often with a water-storing peat layer 1 to 5 m thick [*Sheng et al.*, 2004], whereas wetlands in the tundra generally do not have thick peat layers. Along watercourses, wetlands tend to reduce the variability in river flows, both storing water during high flows and acting as reservoirs of runoff generation during dry summer periods. In the tundra, small lakes sometimes form systems where large amounts of water may accumulate, with high propensity for thermokarst development. When thawing, ice complexes in Yedoma soils form secondary thermokarst objects called *alases* [*Morgenstern et al.*, 2011]. Wetlands and peatlands have high adsorption and ion exchange capacity, which may lead to accumulation of water-transported metals (including heavy metals) and carbon. Wetlands also contribute to retention of pollutants and nutrients, although this depends on the share of river flow that circulates through the wetland [*Quin et al.*, 2015]. The ion composition and suspended material flows differ during summer flows and high flows [*Raymond et al.*, 2007; *Frey and McClelland*, 2009] (see also *Wrona et al.* [2016], for further discussion).

In addition to the regions described above, the riverine coastal domain [see *Carmack et al.*, 2016] could be separately considered as a region with unique hydrodynamic and biogeochemical properties, which may also grow in importance as erosion progresses, permafrost thaws, and sea ice declines. Along the coast, erosion, tidal, and surge processes, as well as water, heat, and geochemical fluxes, are under joint influence of riverine, marine, and atmospheric processes and affect riparian ecosystems and mesoscale coastal landscapes. This emerging research domain is discussed further in *Carmack et al.* [2016] and *Prowse et al.* [2015b].

3. Past Changes and Key Drivers

In this section, we review past changes to the main processes of the Arctic terrestrial freshwater system, with focus on freshwater storages and fluxes. Following the processes, we also highlight past changes in the Arctic hydrophysiographical regions outlined in Figure 1.

3.1. Processes

With regard to precipitation processes, we focus on snow cover changes (see *Vihma et al.* [2016] and *Lique et al.* [2016] for other changes to precipitation). Syntheses of several ground and satellite observational data sets indicates that Northern Hemisphere snow cover extent decreased by 2.2% per decade, averaged for March and April, and by 14.8% per decade for June, over the period 1979–2012 [*Brown and Robinson*, 2011] (updated in *Vaughan et al.* [2013]). Both positive and negative regional trends are distributed throughout the pan-Arctic, however, including spatially distinct areas of increasing and decreasing snow water equivalent (SWE) or snow season length [*Callaghan et al.*, 2011]. In spite of strong regional variability—for example, increasing SWE in northern Eurasia—snow is, by most measures, generally decreasing throughout the Arctic as shown by *Liston and Hiemstra* [2011]. They used a physically based spatially distributed snow modeling system (SnowModel) in conjunction with NASA Modern-Era Retrospective Analysis for Research and Applications atmospheric reanalysis data to show that snow cover onset is later, the snow-free date in spring arrives earlier, and snow cover duration has decreased over the period 1979–2009 [*Liston and Hiemstra*, 2011]. These changes in snowpack are related to the earlier onset of the snowmelt runoff [*Tan et al.*, 2011]. Despite the shorter duration of snow cover, however, hydrological modeling experiments show that an increase in SWE, as observed in the northern parts of the large Eurasian basins, may still lead to increased annual runoff [*Troy et al.*, 2012]. Main drivers of snow changes are air temperature increase and changing amounts and timing of precipitation. The modeled sensitivity of snow to these drivers varies strongly across climates and regions [*Brown and Mote*, 2009]. As winter air temperature increases, observations of rain-on-snow events become more common in Arctic regions where they were rarely seen before [*Ye et al.*, 2008; *Nowak and Hodson*, 2013; see also *Vihma et al.*, 2016; *Instanes et al.*, 2016; *Wrona et al.*, 2016].

In terms of evapotranspiration, it is challenging to determine large-scale changes as observations are scarce. *Zhang et al.* [2009] developed an ET algorithm driven by satellite remote sensing inputs to assess spatial patterns and temporal trends in ET over the pan-Arctic basin 1983 to 2005 and found a mean trend of +0.38 mm yr^{-2} ($p < 0.05$). Negative ET trends occurred over 32% of the region, primarily in the boreal forests of southern and central Canada. *Rawlins et al.* [2010] showed that model trends of annual ET are significantly positive, with a multimodel mean trend (1950–1999) of +0.17 mm yr^{-2} ($p < 0.1$). The differences between the estimates suggest an amplification of ET increases over the recent decades. Increased ET is to be expected with observed warming due to the ability of the atmosphere to hold more moisture at higher temperatures, but ET changes are also driven by landscape alterations, such as vegetation shifts, fires, and lake and reservoir changes [*Wrona et al.*, 2016]. For example, increases in shrubs and trees enhance evapotranspiration and thereby the loss of water from the basin, which leads to drier surface conditions [*Wrona et al.*, 2016].

A large-scale pattern of observed river discharge increases has been frequently reported for Arctic rivers [*Peterson et al.*, 2002, 2006; *McClelland et al.*, 2006; *Shiklomanov and Lammers*, 2009; *Dyurgerov et al.*, 2010; *Overeem and Syvitski*, 2010; *Holmes et al.*, 2013; *Bring and Destouni*, 2014], although decreases are also noted, particularly for some North American rivers [*Déry et al.*, 2005; *McClelland et al.*, 2006]. However, over the last couple of decades, discharge in North American rivers reversed this trend and have increased as well [*Déry et al.*, 2009], aligning them with the more general pattern of increasing discharge. The total river flows, calculated from averages of reanalysis and in situ data, have likely increased from 3900 ± 390 km^3 during 1980–2000 to 4200 ± 420 km^3 during 2000–2010 [*Haine et al.*, 2015], with errors assumed to be about 10%. In addition to average flows, changes in low and high flows have also been reported. *Spence et al.* [2014] have shown that increased late autumn rains may cause the enhancement of winter flow and impact geochemical fluxes from headwater catchments of the subarctic Canadian Shield. *Shiklomanov et al.* [2007] reported that increases and decreases in station maximum daily flows were equally common, with very few significant trends in summer months, which led them to question the generally expected future increase in floods. Increases in station low flows, which are reported for several regions [*L. C. Smith et al.*, 2007; *St. Jacques and Sauchyn*, 2009; *Ehsanzadeh and Adamowski*, 2010; *Rennermalm et al.*, 2010; *Karlsson et al.*, 2012; *Walvoord et al.*, 2012; *Karlsson et al.*, 2015], are consistent with recession flow analyses [*Lyon et al.*, 2009; *Lyon and Destouni*, 2010; *Brutsaert and Hiyama*, 2012] and modeling of permafrost thaw [*Bense et al.*, 2009, 2012; *Frampton et al.*, 2011, 2013] that indicate an increasing contribution of groundwater to streamflow [*Walvoord and Striegl*, 2007]. Overall, a number of reanalysis, modeling, and observation-based studies show that increased atmospheric moisture transport (AMT), due to higher temperatures and changing atmospheric circulation, is likely the principal driver of long-term increases in flow [*Zhang et al.*, 2008, 2013; *Rawlins et al.*, 2009; *Troy et al.*, 2012] (see discussion of atmospheric changes in *Vihma et al.* [2016]). Model-observation studies have also shown that winter precipitation stored as snow is a central component of the AMT contribution [*Troy et al.*, 2012; *Zhang et al.*, 2013]. In contrast, ground ice melt and evaporation from increasingly ice-free ocean water have not been responsible for any greater contribution to historical flow increases [*McClelland et al.*, 2004; *Pavelsky and Smith*, 2006; *Zhang et al.*, 2013], although an exact quantification remains elusive.

Permafrost thaw has changed the hydrological regime in some basins, particularly through altered surface and subsurface interactions. For example, *Connon et al.* [2014] report observed increases of annual runoff in the lower Liard Valley (NWT, Canada) by between 112 and 160 mm over the period of 1996–2012, mainly due to increase of plateau runoff contributing areas and a change in the relative proportions of the major land cover types, such as peat plateaus, channel fens, and flat bogs. These values are large relative to average annual runoff values of between 147 and 216 mm for the period. Changes to temperature and precipitation have also interacted to produce changes in evapotranspiration, runoff, peak seasonal snow accumulation, and snow season length in permafrost basins. Short-term changes in air temperature, ice cover, and soil moisture do not trigger systematic hydrological shifts in permafrost, although they provide a memory that manifests itself during the next warm season, as shown in a model-observation study [*Park et al.*, 2013]. Long-term changes in climate, however, alter the landscape structure and its runoff formation, principally through an increased ALT, as indicated by both observations and modeling [*Quinton et al.*, 2011; *Frampton et al.*, 2013]. In Eurasia, the ALT has generally increased due to thicker snowpacks and high summer soil moisture [*Park et al.*, 2013]. In the Mackenzie and Yukon basin, on the other hand, combined effects of less insulation caused by thinner snow depth and drier soil during summer have partly offset warming effects on the ALT [*Park et al.*, 2013], although these results are based on modeling and further research is needed to confirm them.

Ice cover observations on Northern Hemisphere lakes indicate a shortening duration, with the time of breakup generally changing more rapidly than the freezeup [*Benson et al.*, 2011]. Trends over 1855–2004 were steeper than over 1905–2004, but the most rapid changes occurred in the most recent 30 year period, with freezeup 1.6 d/decade later, breakup 1.9 d/decade earlier, and ice duration 4.3 d/decade shorter. Although ice cover tends to be more sensitive to air temperature variations at lower than at higher latitudes [*Livingstone et al.*, 2010], remote sensing observations indicate that ice cover loss seems to be more rapid in very high latitude lakes [*Latifovic and Pouliot*, 2007]. High-latitude lakes lost ice cover at 1.75 d yr^{-1} during the 1970–2004 period of most rapid depletion, which is more than 4.5 times the rate of lakes in southern Canada. It is unclear whether this reflects the more recent and greater high-latitude warming or potential differences in observational techniques [*Prowse and Brown*, 2010]. Potential effects of reduced lake ice cover on feedbacks to the atmosphere are further discussed in *Vihma et al.* [2016].

In terms of river ice observations, *Beltaos and Prowse* [2009] noted an almost universal trend toward earlier breakup dates but considerable spatial variability in those for freezeup. Changes are often more pronounced during the last few decades of the twentieth century. Overall, twentieth century warming has lead to a 10 to 15 day advance in breakup and delay in freezeup, respectively [see also *Lique et al.*, 2016]. This is in agreement with with earlier estimates [*Magnuson et al.*, 2000], but the relationship is complicated by changes in snow accumulation and spring runoff [*Beltaos and Prowse*, 2009].

The discharge from large Arctic rivers (examples from Eurasian rivers shown in Figure 3) integrates all features and change drivers from the hydrophysiographic regions that make up their catchments [*Holmes et al.*, 2013]. The seasonal character of discharge, which for many large Arctic rives is affected by dams [*McClelland et al.*, 2004; *Yang et al.*, 2014], shows quite high flow during all of the summer period and very low flows in winter (Figure 3a). Interannual variability is high, and the total water flow in a wet water year can be twice the flow in a dry year (Figure 3b). As noted above, there are examples of both increases and decreases in river discharge to the Arctic Ocean (Figure 3c), but for the Arctic as a whole, flow has increased both from Eurasia and North America when considering the longest possible time periods (NOAA Arctic report card 2011, available at www.arctic.noaa.gov/reportcard [*Holmes et al.*, 2013]).

3.2. Hydrophysiographic Regions

In this section, past changes and key drivers are summarized for each hydrophysiographic region (Figure 1). Annual river flows observed in the tundra region have generally increased, but there are exceptions [*Zhang et al.*, 2009; *Overeem and Syvitski*, 2010]. Increasing temperatures, and to a large degree, precipitation, are the main drivers of change in Arctic tundra hydrology [see *Vihma et al.*, 2016; *Lique et al.*, 2016]. The warming effect in sheltered locations has created polar oases in the otherwise barren high Arctic [*Edlund and Alt*, 1989], and remote sensing shows that recent warming has led to increased development of thermokarst lakes in ice-rich permafrost environments [*Smith et al.*, 2005]. Consecutive positive observed anomalies of snow depth and rainfall have likely contributed to the widespread warming of near-surface permafrost in the central Lena River basin [*Iijima et al.*, 2010]. Satellite estimates of evapotranspiration in the tundra show increases by 2.7–2.8 mm/decade over 1983–2005, but this figure is small in relation to annual values and only significant at $p = 0.1$ for North America, and not significant for Eurasia [*Zhang et al.*, 2009; see also *Vihma et al.*, 2016].

Shrub expansion has occurred throughout the Arctic tundra during the past 50 years [*Sturm et al.*, 2001, 2005; *Tape et al.*, 2006, 2012], modifying the microclimate of the landscape and altering erosion and biogeochemical fluxes. The underlying mechanisms of shrubification are largely attributed to increasing air and soil temperature, but factors related to local hydrological changes such as precipitation, soil moisture, and snowpack also influence shrub growth [*Wrona et al.*, 2016].

While there is evidence of northward advance of the tree line across the boreal plains, with increase in tree density and canopy size of individual trees [*Danby and Hik*, 2007], other parts of the boreal forest are experiencing degradation of their discontinuous permafrost substrate, which has reduced the occurrence of tall-growing vegetation [*Quinton et al.*, 2011]. Large-scale satellite-based estimates indicate that evapotranspiration has increased in Eurasian forests (6.94 mm/decade, $p < 0.05$) but that it may have decreased in North American ones (−3.06 mm/decade, no significance) [*Zhang et al.*, 2009]. River flows have generally increased in the boreal plains and shield regions, but decreases are reported for some Canadian rivers [*Zhang et al.*, 2009; *Déry et al.*, 2011].

In shield regions, extensive hydropower development [see *Instanes et al.*, 2016] has also modified the flow of many rivers, lowering summer flows and increasing winter flows [*Déry et al.*, 2011]. Information on undeveloped rivers, with flow patterns unaltered by dam construction, is relatively scarce in the Arctic shield regions. For the Tana River in northern Norway and Finland, long-term station records do not reveal any significant trends in streamflow but indicate increases in both annual precipitation and annual precipitation variability [*Dankers*, 2003]. In contrast, streamflow variability has decreased in unregulated rivers on the Canadian Shield that drain into Hudson Bay [*Déry et al.*, 2011].

In some mountainous regions, thawing of alpine permafrost has modified the pathway, timing, and amount of runoff. Using measured data and reinforced by modeling based on *Frampton et al.* [2013], *Sjöberg et al.* [2013] found that accompanying observed permafrost degradation in northern Sweden, there are higher recession flows and increased winter discharge sustained by groundwater in most alpine basins they studied, although these signals were not consistent across landscapes and, for minimum flows, not correlated strongly with permafrost extent. Mountainous regions also contain intermontane basins. Compared with their surrounding heights, these basins have lower elevations and gentler topography and show earlier response to climatic change. In the Yukon Flats in Alaska, there are indications from remote sensing that expanded shallow suprapermafrost or intrapermafrost taliks are linked to the expansion and shrinking of lakes [*Jepsen et al.*, 2013].

On the low-latitude grasslands in the upstream portion of the largest Arctic river basins, reservoirs of different sizes have been constructed for irrigation, increasing evaporation. Groundwater has also been exploited to water crops. In the upper Ob, water flows have decreased in summer and increased in winter, as a result of increasing water abstraction for agriculture and industry and flow regulation from dams [*Yang et al.*, 2004a]. Similar effects have been observed in the upper Yenisey basin [*Yang et al.*, 2004b].

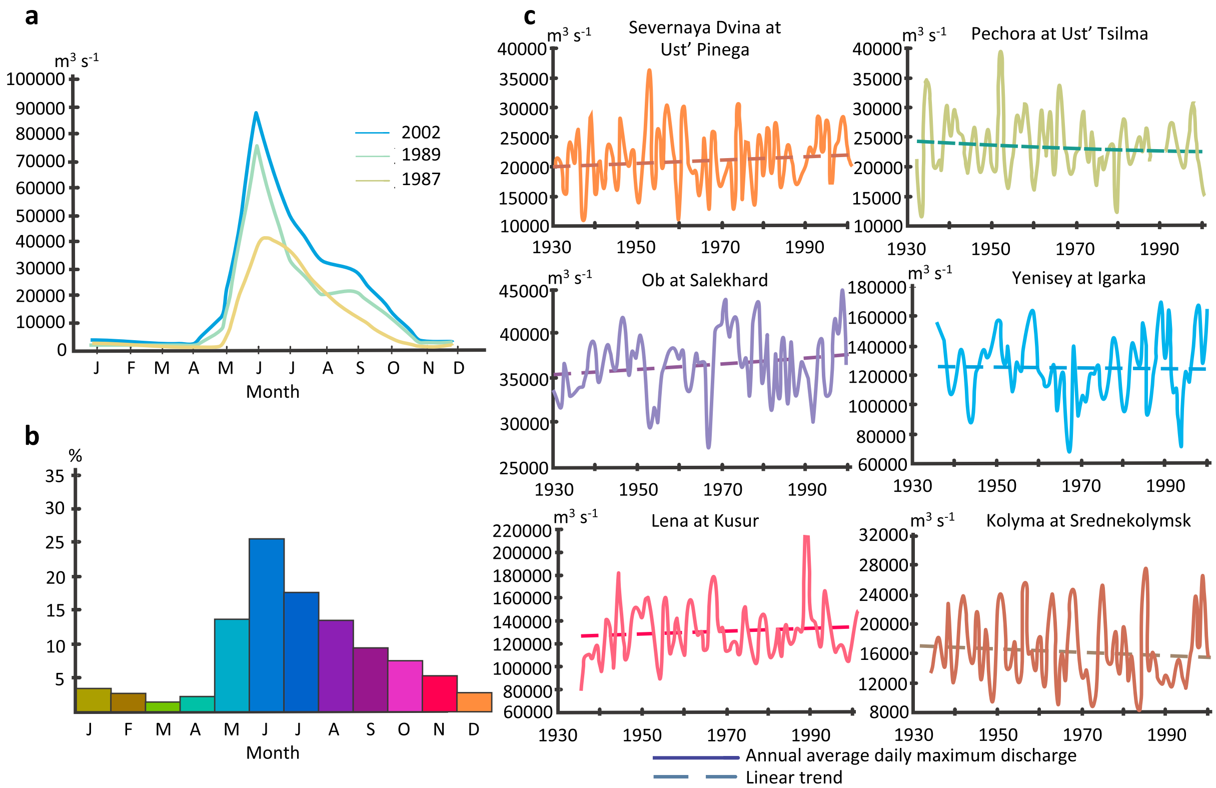

Figure 3. Examples of discharge patterns from large Eurasian Arctic rivers. (a) Monthly average discharge for the combined Eurasian Arctic drainage. (b) Differences in discharge between an average, a dry, and a wet year for the Lena River. (c) The long-term annual river discharge for the six largest Eurasian Arctic rivers from 1936 to 2002. Due to the declining accuracy of gauge readings in Russia [*McClelland et al*., 2015], we only include data until 2002 in Figure 3. Even considering gauge uncertainty in the period since then, flows distinctively reached a record high in 2007 [*Rawlins et al*., 2009; *Shiklomanov and Lammers*, 2009], and the years from 2000 to at least 2009 have been wetter than the average for the period 1950–2009 [*Walsh et al*., 2011]. All river data are available from the Global Runoff Data Centre (http://www.bafg.de) and the R-ArcticNET database (http://www.r-arcticnet.sr.unh.edu).

Of the several thousands of glaciers that are located in the pan-Arctic, only ~25–30 have ongoing operational mass balance programs to quantify glacier mass balance conditions and change [*World Glacier Monitoring Service*, 2013] (www.wgms.ch). Analyses of glacier area fluctuations, based on historical accounts, aerial photography, and satellite images, are more numerous. Since the early 1990s, the increasing glacier [e.g., *Kaser et al*., 2006; *Cogley*, 2012; *Vaughan et al*., 2013; *Mernild et al*., 2014] and Greenland ice sheet (GrIS) net mass loss and surface runoff have followed atmospheric warming [e.g., *Hanna et al*., 2008; *Box and Colgan*, 2013; *Church et al*., 2013; see also *Carmack et al*., 2016; *Lique et al*., 2016]. Mass loss from the GrIS has increased rapidly, with recent estimates of $375 \pm 24\ km^3\ yr^{-1}$ for 2011–2014 [*Helm et al*., 2014] (based on estimations from CryoSat-2) and $575 \pm 95\ km^3\ yr^{-1}$ for September 2011 to September 2012 (NOAA Arctic report card 2012, available at www.arctic.noaa.gov/reportcard and based on Gravity Recovery and Climate Experiment, a satellite mission used to detect changes in water mass). For Greenland in its entirety, runoff was recently estimated to $481 \pm 85\ km^3\ yr^{-1}$ for 1960–2010 [*Mernild and Liston*, 2012] (based on modeling). On the pan-Arctic scale, the cumulative glacier net mass balance has been negative for all glacier regions for the period 1979–2009 [*Mernild et al*., 2014], and overall contributions to sea level change are positive (see glacier change summary in Table 1). For some smaller basins, mass balance change constitutes a substantial share of river runoff. For example, about 35% of runoff from the Mittivakkat Glacier basin in southeast Greenland originates from net mass balance loss [*Liston and Mernild*, 2012].

Studies of past changes to wetlands indicate both increases and decreases in areal extent. Generally, the changes in number and area of wetland lakes and ponds depend on the local permafrost and hydrological conditions. Lakes have mostly decreased in size and number in areas where discontinuous and sporadic permafrost thaw is in progress [*Andresen and Lougheed*, 2015]. In these areas, flow paths and water tables are becoming deeper, and lakes may also drain as the permafrost substrate is breached. However, other factors,

Table 1. Regional Examples of Glacier Area Changes and Sea Level Contributions in the Pan-Arctic Drainage Basin

Glacial Area Changes			
Region	Time Period	Change in Area	Reference
Yukon Territory	1958/1960–2006/2008	−22%	*Barrand and Sharp* [2010]
British Columbia	1985–2005	−11%	*Bolch et al.* [2010]
Interior northern Baffin Island	1958–2005	−55%	*Anderson et al.* [2008]
Southeast Baffin Island	1920–2000	−13%	*Paul and Svoboda* [2010]
Southeast Greenland	1986–2011	−27%	*Mernild et al.* [2012]
Northern Polar Urals	1953/1960–2000	−22%	*Shahgedanova et al.* [2012]
Suntar Khayata Region, North Asia	1945–2002/2003	−19%	*Ananicheva et al.* [2006]
Koryak Upland near Kamchatka Peninsula	1950s–2003	−67%	*Ananicheva and Kapustin* [2010]
Sea Level Contributions			
Region	Time Period	Contribution to Sea Level Equivalent (SLE) in mm yr^{-1}	Reference
Alaska	1999–2009	0.13 ± 0.05	*Mernild et al.* [2014]
Arctic Canada (North)	1999–2009	0.10 ± 0.01	*Mernild et al.* [2014]
Greenland (nonice sheet)	1999–2009	0.09 ± 0.01	*Mernild et al.* [2014]

such as changes in temperature and evapotranspiration, are also driving forces [*Karlsson et al.*, 2015]. In contrast, increases, decreases, or no change in lake size and number are reported from areas in continuous permafrost [*Jones et al.*, 2011; *Andresen and Lougheed*, 2015].

4. Projected Changes and Key Drivers

In this section, we review main projected changes to the central processes of the Arctic terrestrial freshwater system, with focus on freshwater storages and fluxes. Following the processes, we also highlight projected hydrological changes across the Arctic hydrophysiographical regions outlined in Figure 1.

4.1. Processes

In Table 2, we synthesize reported projections for hydrological processes across the hydrophysiographical regions. Projected changes in seasonal snowfall and snow water equivalent are spatially variable and depend on local climate conditions. In very cold regions, increased winter precipitation will lead to a deeper snow cover (and higher SWE), while in warmer regions, higher temperatures will lead to the opposite [*Räisänen*, 2008; see also *Lique et al.*, 2016]. However, other snow-related variables, such as snow cover extent (SCE), exhibit a more direct relationship with air temperature.

Brutel-Vuilmet et al. [2013] found a strong negative correlation between Northern Hemisphere spring SCE and the corresponding mean surface air temperature, and by 2080–2099, the average reduction of seasonal SCE varies from $7.2 \pm 3.8\%$ for RCP2.6 to $24.7 \pm 7.4\%$ for RCP8.5, relative to a 1986–2005 reference period. Other modeling studies have also showed increases in snowfall and SWE associated with the projected increase in cold season precipitation in northeastern Eurasia and northern Canada, while they decreased in more southerly places in which warming effects dominated [*Räisänen*, 2008; *Deser et al.*, 2010; *Krasting et al.*, 2013]. In these studies, the −10°C and −20°C winter air temperature isotherm lines represent transition boundaries for snowfall [*Deser et al.*, 2010; *Krasting et al.*, 2013] and SWE [*Räisänen*, 2008], respectively.

There is growing evidence for an increase in precipitation extremes, with relative increases generally exceeding those for annual mean precipitation under the projected 21st century global warming [*Kharin et al.*, 2013; *Sillmann et al.*, 2013; *Vihma et al.*, 2016]. A statistically significant signal of increasing summer season precipitation across the Arctic as a whole is likely to emerge from the variability when global temperature rise surpasses 1.4°C, which is likely to occur around 2040 [*Mahlstein et al.*, 2012]. Signals in extreme precipitation, however, as well as regional signals, may emerge earlier or later, and because models underestimate past changes, the time of emergence for the entire Arctic is likely too conservative [*Mahlstein et al.*, 2012].

A diagnostic study of future evapotranspiration changes projected in CMIP5 climate models under the RCP45 scenario by *Laîné et al.* [2014] shows a change on the order of 0.05 mm d^{-1} in winter and 0.25 mm d^{-1} in summer over a 100 year period for the terrestrial pan-Arctic (1980–2000 versus 2080–2100) [see also *Lique et al.*, 2016;

Table 2. Projected Changes to Hydrological Processes Across Arctic Hydrophysiographical Regions

Process	Hydrophysiographical Region	Projected Changes	Specific Effects/Comments
Precipitation	Arctic tundra	Generally increasing annual precipitation (highest increase in winter and fall)	Increase in snow water equivalent, earlier snowmelt, decrease in seasonal snow cover period, warming of permafrost, and increase in coastal erosion
	Boreal plains	Generally increasing annual precipitation (highest increase in winter and fall), with some regional decreases in summer precipitation	Increase in soil moisture and runoff, midwinter snowmelt events and more frequent rain on snow events, possible soil moisture decrease (drying) in summer, decrease in maximum snow water equivalent, and decrease in seasonal snow cover extent in May/June
	Mountains	Generally increasing annual precipitation (highest increase in winter and fall), with some regional decreases in summer precipitation	Increase in soil moisture and runoff and decrease in seasonal snow cover extent in May/June
	Shields	Generally increasing annual precipitation (highest increase in winter and fall), with some regional decreases in summer precipitation	Increase in soil moisture and runoff, midwinter snowmelt events and more frequent rain on snow events, and possible soil moisture decrease (drying) in summer
	Ice sheets, ice caps, and glaciers	Increasing annual precipitation	Increase in snowpack/accumulation
	Wetlands	Increasing annual precipitation (highest increase in winter and fall)	Midwinter snowmelt events and more frequent rain on snow events and decrease in seasonal snow cover extent in May/June
	Grasslands	Small increase in annual precipitation (with decreases in summer precipitation)	Decreases in summer precipitation with possible increase in summer drought events
Evapotranspiration	Arctic tundra	Increasing annual evapotranspiration (more in summer than winter)	Vegetation and ecosystem changes
	Boreal plains	Increasing annual evapotranspiration (more in summer than winter)	Possible drying (decrease in soil moisture) in summer
	Mountains	Increasing annual evapotranspiration (more in summer than winter)	Possible drying (decrease in soil moisture) in summer
	Shields	Increasing annual evapotranspiration (more in summer than winter)	Possible drying (decrease in soil moisture) in summer
	Ice sheets, ice caps, and glaciers	Increasing annual evapotranspiration and sublimation	Increasing surface mass loss
	Wetlands	Increasing annual evapotranspiration (more in summer than winter)	Possible decrease in wetland area
	Grasslands	Increasing annual evapotranspiration (possible decrease in summer)	Possible decrease in summer evaporation due to decrease in summer precipitation
Runoff and river flux	Arctic tundra	General increase in mean, high, and low flows and limited areas of decrease in high and low flows	Possible high and low flow decreases in western Siberia
	Boreal plains	General increase in mean flows but limited areas of decrease and variable changes in high and low flows	Possible high and low flow decreases in western Siberia, southern Canada
	Mountains	Consistent increase in mean flows, high and low flows generally increasing	Possible low flow decreases in western Siberia, southern Canada
	Shields	Smaller increase in mean flows and variable changes in high and low flows	Possible low flow decreases in southern Canada and Scandinavia
	Ice sheets, ice caps, and glaciers	Increase in runoff as glacier mass loss increases, followed by eventual runoff decrease as glacier shrinkage counteracts the increased melting	Initially increasing surface ablation/mass loss, thereafter a drop in surface ablation following glacier shrinkage
	Wetlands	General increase in mean flows and variable changes in high and low flows	Possible high and low flow decreases in western Siberia and western Canada
	Grasslands	Variable change in mean, low, and high flows	Increases more likely in Asia and decreases more likely in North America
Permafrost and groundwater hydrology	Arctic tundra	Increases in active layer depth and additional degradation of permafrost will provide greater reservoirs for	Soil moisture increasing on plateaus. Vegetation changes will increase evapotranspiration. Succession of

Table 2. (continued)

Process	Hydrophysiographical Region	Projected Changes	Specific Effects/Comments
		subsurface storage of groundwater and near-surface soil moisture. General increase in low flows due to strengthened connections between surface water and groundwater and increase in runoff and erosion.	wetting and drying of landscapes. Suprapermafrost groundwater will increase, providing a limited reservoir for later season flow to streams and rivers.
	Boreal plains	Complete disappearance of permafrost in marginal areas of sporadic occurrence. General increase in low flows due to strengthened connections between surface water and groundwater.	Where groundwater table is below the surface water, infiltration may leave the surface drier. In areas of groundwater upwelling, increased connectivity may support wetland development.
	Mountains	Limited change in groundwater dynamics but possible increases in low flows for valley and wetland regions in areas of degrading permafrost	In regions of discontinuous permafrost, springs tend to be found on slopes where percolating water first encounters permafrost and water is forced to the surface. Location of springs may shift.
	Shields	Limited change in groundwater dynamics. In regions of increasing summer aridity, possible increasing seasonal loss of connectivity between lakes, wetlands, and rivers.	Zones of recharge and discharge over thick shields are mostly tied to fractures and will generally remain unchanged.
	Ice sheets, ice caps, and glaciers	Limited change in groundwater dynamics; however, increase in groundwater storage is possible.	Possible increase in groundwater storage if the increased rate of ice melt is greater than the rate of groundwater discharge. If groundwater reservoir already saturated (as is typical), ice melt will emerge as overland flow with no change in groundwater.
	Wetlands	Increasing inflow to wetlands/peatlands and small lakes which may cause them to expand. Variable change in moisture content of wetland and peatland areas. Changes to water and energy balance. Draining of thermokarst lakes.	As permafrost degrades, infiltration to groundwater will increase and surface runoff will decrease. These changes will increase groundwater storage and flow, yielding greater discharge to wetlands in polar regions. Flushing of preserved contaminants. Secondary thermokarst lakes forming.
	Grasslands	Groundwater will be deep enough to permit little upward connectivity. Groundwater/surface water interactions dominated by infiltration.	As permafrost degrades and surface water is no longer held near the surface, the groundwater table will recede to depths greater than boreal tree species can reach, increasing the occurrence of ecosystems appropriate for grasslands.
River and lake ice	Arctic tundra	River and lake ice will continue to form every winter. May be thinner in rivers that start to maintain continuous flow through the winter. Lake ice may be thinner as winter snowfall increases.	With greater snowpack and insulation from the cold, ice thickness will decrease.
	Boreal plains	Ice thickness in ponds likely to decrease in total thickness. Ice cover in rivers and streams likely to become thinner with more frequent occurrences of midwinter open water.	Ice thickness likely to decrease due to longer periods of groundwater input, greater total input, and increased snow depths.
	Mountains	Ice on mountain lakes likely to decrease in thickness and duration. Ice changes on streams dependent upon terrain and source.	Air temperatures and snow depth will decrease end of winter lake ice thickness, but terrain controls on mountain streams ice will likely not change. Winter icings (or aufeis) could increase markedly if sources of groundwater flow later into the winter.
	Shields	Ice on lakes in shield region is likely to decrease in thickness and duration. Ice on rivers and streams will decrease in	Thicker snowpack, milder temperatures, and greater input of potentially warmer groundwater will contribute to decreased ice.

Table 2. (continued)

Process	Hydrophysiographical Region	Projected Changes	Specific Effects/Comments
		thickness with more frequent midwinter open water.	
	Ice sheets, ice caps, and glaciers	NA	
	Wetlands	Ice on lakes in wetlands is likely to decrease in thickness and duration. Ice on rivers and streams will decrease in thickness with more frequent midwinter open water.	Thicker snowpack, milder temperatures, and greater input of potentially warmer groundwater will contribute to decreased ice.
	Grasslands	Ice on lakes in grasslands is likely to decrease in thickness and duration. Ice on rivers and streams will decrease in thickness with more frequent midwinter open water.	Thicker snowpack, milder temperatures, and greater input of potentially warmer groundwater will contribute to decreased ice.

Vihma et al., 2016]. Based on suite of nine general circulation model (GCM) simulations from CMIP3 over the terrestrial pan-Arctic, *Rawlins et al*. [2010] showed that over the 100 year period from 1950 to 2049, annual evapotranspiration trends range from 0.24 mm yr^{-2} to as much as 0.92 mm yr^{-2}, with the multimodel mean trend at 0.65 mm yr^{-2}. In general, results suggest acceleration in evapotranspiration over the latter half of the present century.

GCMs generally indicate higher water flows in the future [*Kattsov et al*., 2005, 2007; *Holland et al*., 2007; *Rawlins et al*., 2010] (see also detailed discussion about climate models in *Lique et al*. [2016]). Analyses using a macroscale hydrological model with forcing data from GCMs show projected increases in discharge of up to 31% by the 2080s [*Arnell*, 2005]. Simulations with a global hydrological model indicate consistent increases of 25–50% across forcing data from three GCMs over the most of the pan-Arctic, except in the lower Ob and Hudson Bay drainages where some decreases are projected [*van Vliet et al*., 2013]. Similar results were obtained with a larger set of GCM forcing data in a study by *Koirala et al*. [2014]. Model consistency is generally higher for increases than decreases in the Arctic. In contrast to historical flows discussed in section 3.1 above, simulations indicate that future flows will be increasingly driven by evaporation increases from retreating ice cover in the Arctic [*Bintanja and Selten*, 2014] (see also discussion in *Vihma et al*. [2016]). Projections for high and low flows (expressed as 95th and 10th percentiles of daily values) diverge, with increases of 25–50% consistently projected for eastern Siberia and high Arctic North America and decreases of 0–25% mostly projected for western Siberia, lower latitude North American drainage, and for high flows, also Scandinavia [*van Vliet et al*., 2013]. However, as the climate forcing is the most important control on the hydrological model output [*Arnell*, 2005], and as climate models do generally not represent permafrost dynamics adequately [*Koven et al*., 2013; *Slater and Lawrence*, 2013], further studies would be needed to confirm these results—for example, results of *Koirala et al*. [2014] indicate a greater occurrence of increases in low flows. Furthermore, controls on minimum flows are complex and scale dependent [*Rennermalm et al*., 2012], so the low flow results are likely not true for basins of all sizes in the described regions. In general, however, a tendency of increasing river flows is likely, which will raise the transport rates of nutrients, sediment, and carbon in Arctic rivers. Older carbon will also be increasingly mobilized [*Aiken et al*., 2014]. In addition, north-south gradients of flow availability may change, increasing pressure on water resources in southern basins (see further discussion in *Instanes et al*. [2016] and *Vihma et al*. [2016]).

Irrespective of emission scenario, Arctic permafrost area is projected to decline at a semilinear rate until the 2040s, when decline decelerates in lower emission scenarios. In a high-emission scenario, decline continues at approximately the same rate until 2100 [*Slater and Lawrence*, 2013; see also *Lique et al*., 2016]. The ongoing permafrost thaw triggers hydrological regime changes in many ways. With a thickening active layer, some areas that currently store water will change into ones that produce runoff, but trends are different for various landscapes. Furthermore, warming can affect soils in opposing directions, depending on the hydrologic gradient: additional meltwater may cause soil humidification, but it could also be the initial step toward soil drying due to erosion and streamflow intensification. Wetting is more likely in lowlands and plateaus and drying more likely on steeper slopes, where erosion may also increase. Apart from warming, precipitation also influences permafrost, with high prewinter rainfall and snowfall accelerating soil warming through greater latent heat of freezing and greater snow insulation [*Iijima et al*., 2010]. The temperature of rain is also important, and as rain becomes more frequent, its higher heat content in comparison with snow will influence the melting of snow and the active layer thickness. *Walvoord et al*. [2012] showed that diminishing permafrost increases the spatial extent of groundwater discharge in lowlands and decreases the proportion of suprapermafrost (shallow) groundwater contribution to total base flow.

Lake ice cover will decrease, according to a recent modeling study, in both thickness (−10 to −50 cm) and duration (−15 to −50 days) during 2040–2079 compared to 1960–1999 [*Dibike et al*., 2011]. The largest changes are projected for the Pacific coastal regions of North America, northeastern Canada, eastern Europe, Scandinavia, and northern Russia. The snow depth on lake ice is projected to change by −20 to +10 cm and the amount of white ice (i.e., ice forming from wet snow on top of ice) by −20 to +5 cm. In the high latitudes, white ice may form more easily in the future due to increasing snowfall and thinner ice cover [see discussion of changing lake ice effects in *Instanes et al*., 2016; *Vihma et al*., 2016; and *Wrona et al*., 2016].

A future reduction in thermal gradients along northward flowing and ice-covered Arctic rivers has been suggested to decrease spring flooding because of lessening in the severity of ice jamming [*Prowse et al*., 2010]. On the other hand, snow water equivalent in spring is projected to increase at very high latitudes, particularly in areas with winter temperatures below −30°C [*Adam et al*., 2009]. The net result of these two factors (magnitude of spring snow water equivalent and severity of ice jams) remains to be quantified but will vary by river basin according to spatial and temporal variability in future precipitation accumulation and snowmelt regimes around the circumpolar north, including in the headwaters of the large basins located in more southerly latitudes [*Prowse et al*., 2011]. The effects of changes in river ice breakup regimes on freshwater ecosystems and northern infrastructure are reviewed in *Wrona et al.* [2016] and *Instanes et al*. [2016], respectively.

4.2. Hydrophysiographic Regions

Across the tundra, climate warming will generally deepen the active layer and induce a loss of ground ice. Depending on the ice content, places with ice-rich permafrost will undergo subsidence, which leads to thermokarst and small lake formation that affect surface storage and evaporation [*Smith et al*., 2005; *Jorgenson et al*., 2008]. A thickened active layer will accommodate more groundwater at the expense of surface runoff, and the magnitude of spring floods may decrease. Time is an important factor, as landscapes will transition from initial wetting due to subsidence to later drying from more complete permafrost degradation and talik formation, depending on the local hydraulic gradient. Snow depths are generally projected to increase due to increasing winter precipitation [*Callaghan et al*., 2011; see also *Vihma et al*., 2016]. The southern margins of Arctic tundra will continue to undergo change to shrubs [*Sturm et al*., 2001], with attendant effects on snow catch, melt, evaporation, and herbivores' access to food, shelter and migration, and shifts in ecosystems. In addition, overland transportation, e.g., snow and ice roads, may be affected with increasing hindrance from shrub growth and shorter snow cover period.

In the boreal plains, warming will also deepen the active layers in permafrost regions, and lateral attrition will take place along the edges of permafrost bodies. In some areas, drying of lakes and wetlands will lead to a browning of forests, a decline in tree growth, and a reduction in latent heat flux, although effects of scale and hydraulic gradient will also influence this process. There is evidence of recent browning of the boreal forests in Central Alaska, which may be related to increased drought stress, though the infestation of insects can also be a factor [*Parent and Verbyla*, 2010; *Verbyla*, 2011]. On the other hand, deepening and moistening of the active layer may also create perennially waterlogged conditions that suppress forest development, something that has been observed in the Lena River delta [*Iijima et al*., 2014]. Increasing connectivity between surface water and groundwater may support wetland development in areas of groundwater upwelling. Additional implications of increased connectivity include impacts on erosion, sediment, and nutrient release, processes that in turn affect downstream deltas [*Rachold et al*., 1996, 2000; *Syvitski et al*., 2005; *Frey and McClelland*, 2009; *Rowland et al*., 2010] and form new riparian hydrological regimes. Snowmelt timing change will affect plant growth patterns and through them also evapotranspiration and soil moisture. Permafrost degradation may regionally be accelerated by anthropogenic reductions or removal of insulating peat layers. In the southern parts of the boreal plains, human water demand is likely to increase both for natural resource extraction, such as tar sands development and fracking, and agricultural and industrial use [see *Instanes et al*., 2016]. Fires are likely to increase in frequency [*Stocks et al*., 1998; *Flannigan et al*., 2005; *Hu et al*., 2015]. Possible hydrological effects, either from fires or managed clear-cutting, include higher snow accumulation and subsequent higher volumes of spring meltwater, but such effects are not consistently observed [*Ellis et al*., 2013; *Semenova et al*., 2015].

Bedrock terrain can shed snowmeltwater effectively, and the reduced summer rainfall will leave the shields with diminished moisture available for evaporation. However, climate warming can lead to higher evapotranspiration and eutrophication. Depending on the proportion of exposed bedrock and the coverage of boreal vegetation, the rates are spatially variable. With an increase in aridity in parts of the sub-Arctic, a loss of connectivity between the lakes, wetlands, and rivers will be exacerbated in some regions. This has significant implications for the seasonal reduction or loss of flow, with attendant effects on water quality and ecology. Zones of groundwater recharge and discharge, however, depend mostly on the fracture networks in the bedrock and will not change.

In the mountains, the existence of permafrost in subalpine zones reduces storage and enhances runoff generation, but the shrinkage of alpine permafrost can alter this situation. Climate warming and a shift in the rain-snow regime will affect the streamflow regime. A reduction of permafrost in mountain basins will affect components of the water balance of river basins. It has long been recognized that small basins with a low percentage of permafrost produce lower rainfall peaks but higher recession flows than basins underlain by larger amounts of permafrost [*Slaughter et al.*, 1983]. In a changing climate, amplified warming with elevation [*Mountain Research Initiative EDW Working Group*, 2015] may reinforce changes in Arctic mountain environments.

For grasslands in Canada, one study of a small watershed showed no projected change in annual evapotranspiration and water yield from 1971–2000 to 2041–2070 but a decrease in storage [*Zhang et al.*, 2011]. However, seasonal changes were not consistent, with uncertainties stemming from the choice of climate model. In general, future warming and continuing growth of water abstractions will act to exacerbate water shortage in the grasslands used for agriculture.

Glaciers in the pan-Arctic will mostly continue to thin and retreat, even without additional warming, as they are generally not in balance with the current climate. Regional differences and uncertainties are large, with mean glacier losses between $66 \pm 51\%$ and $11 \pm 28\%$ required to return to equilibrium [*Mernild et al.*, 2014]. A continued lengthening of the ice melt period and possibly warmer summers will raise the elevation of the equilibrium line of glaciers and increase the rate of mass loss. For the pan-Arctic, in general, water fluxes from glaciers will therefore first increase over the near term but will decline as the size of glaciers continue to diminish [*Arctic Monitoring and Assessment Program*, 2012].

In wetlands, a deepening of the active layer may initially lead to increased storage capacity for additional water and to the formation of small lakes. In the long term, however, continued permafrost degradation may lead to drying out of some wetlands, depending on permafrost conditions and hydraulic gradient. Runoff patterns will change erosion rates and possibly give rise to flushing of stored contaminants. The changing water balance of wetlands and peatlands will also influence the energy balance and the hydrochemical characteristics of catchments.

5. Cross-System Effects

Many of the changes outlined in sections 3 and 4 have effects that go beyond terrestrial hydrology to also influence other components of the Arctic freshwater system. In this section, we describe a number of such links, some of which are also schematically highlighted in Figure 4. We stress that Figure 4 is not intended to thoroughly describe these links, which are complex and often bidirectional, but rather to present an overarching view of some points of cross-system interactions. Companion figures that highlight overarching links in a similar way for other components of the AFS are available in *Carmack et al*. [2016], *Instanes et al*. [2016], *Vihma et al*. [2016], and *Wrona et al.* [2016]. Some examples of more detailed and elaborate system diagrams related to freshwater in the Arctic can be found in *Francis et al*. [2009] and *Karlsson et al*. [2011].

5.1. Processes

The broad increase in winter precipitation over the terrestrial Arctic (with regional summer decreases) may have contributed to the overall increase in the aggregate discharge into the Arctic Ocean [*Troy et al.*, 2012]. At the same time, modeling simulations indicate that a small decrease in summer precipitation along with increased potential for evapotranspiration due to the warming climate has resulted in reduced summer soil moisture in various regions of the terrestrial Arctic [*Dirmeyer et al.*, 2013], with cascading effects for structure and function of ecosystems [see *Wrona et al.*, 2016]. The insulating effect of increased snowpack resulting from increased winter precipitation in the high Arctic may result in permafrost warming and increase in

Figure 4. Links between terrestrial hydrology and other components of the Arctic freshwater system. Links are colored depending on whether they are primarily pertaining to permafrost degradation (orange), decrease in snow and ice cover (blue), more surface water flow (purple), or drier or wetter soils (gray). This graphic is neither meant to be comprehensive nor detailed but is intended to illustrate, at a very overarching level, the multifaceted nature of change in the Arctic freshwater system.

active layer thickness, also with effects on ecosystems and resources [see *Instanes et al.*, 2016]. Moreover, the decrease in the proportion of snowfall (versus rainfall) in the low Arctic and the corresponding increase in snowfall in the high Arctic, as well as the decrease in snow cover duration, affect the seasonality of runoff and magnitude of peak flows in Arctic draining rivers, which also has effects on ecosystems, atmospheric feedbacks, and resources.

Evapotranspiration is the largest vertical freshwater flux out of the Arctic drainage basins and therefore the main link from river basins to the atmosphere [see also *Vihma et al.*, 2016]. Through local evapotranspiration-precipitation recycling, it also constitutes a link back to surface waters, a component that may change with both water reservoir construction and irrigation [*van der Ent et al.*, 2010; *Jarsjö et al.*, 2012; see also *Instanes et al.*, 2016] and soil moisture change [*Hinzman et al.*, 2013]. The latter process is also associated with ecosystem changes [*Wrona et al.*, 2016]. Shifts in vegetation type may increase evapotranspiration by 1–13% until the 2050s across the pan-Arctic [*Pearson et al.*, 2013]. Even without vegetation shifts, increases in evapotranspiration through increasing lake areas may contribute to some observed discrepancies between atmospheric and land surface trends [*Karlsson et al.*, 2015].

River flows strongly affect ecosystems [*Wrona et al.*, 2016] and resources [*Instanes et al.*, 2016], in addition to their flux to the Arctic Ocean, which has implications for thermohaline circulation and marine ecosystems [see *Carmack et al.*, 2016]. Deltas, lakes, and wetlands are partly fed by runoff, which will govern water levels and impact aquatic and terrestrial life. Hydropower is fundamentally dependent on the supply of freshwater through rivers but also gives rise to considerable seasonal shifts in discharge. The variability of flow, as well as high and low flows, affects both ecosystems [*Wrona et al.*, 2016] and resources [*Instanes et al.*, 2016], including the use of liquid freshwater as drinking water for the Arctic population [*White et al.*, 2007; *Alessa et al.*, 2008; *Nilsson et al.*, 2013; *Bring et al.*, 2015b].

Large-scale atmospheric patterns also influence flow. More intense, positive phases of the Pacific Decadal Oscillation have been shown in observations to be a factor in above-average winter precipitation in the Alaska coast range but below average in interior Alaska, Canada, and eastern Siberia, and subsequently, decreased river discharge in northern Canada [*Mantua and Hare*, 2002; *Déry and Wood*, 2005].

Changes in permafrost hydrology are coupled to effects on soil moisture, evapotranspiration, flow pathways, and storage, both in groundwater and surface water, and in ground ice. Where permafrost changes are widespread across the landscape, hydrological effects include modified exchanges of moisture with the atmosphere, lake changes that affect ecosystems, changes in stability and flow patterns that affect resources, and biogeochemical fluxes and erosion that affect the ocean.

A reduction in river and lake ice will influence transportation opportunities in cold and remote regions [see *Instanes et al.*, 2016]. Given that ice-covered freshwater bodies comprise a significant portion of the high-latitude, sub-Arctic and tundra landscape, they need to be considered in climate modeling of this region [*Prowse et al.*, 2011; *Lique et al.*, 2016; *Vihma et al.*, 2016].

5.2. Hydrophysiographic Regions

The Arctic tundra has a direct connection to the riverine coastal domain that encircles the Arctic Ocean. Many small to medium size rivers that originate in the tundra flow directly to the ocean, but the flow magnitude of most of them remains unknown. In the high Arctic, ongoing resource development concentrated along the coast [*Andreeva*, 1998] may give rise to large local effects on hydrology. Over tundra areas, in general, the exchange with the atmosphere through the boundary layer energy balance is strongly affected by snow and the surface albedo of wet or dry tundra.

In boreal and shield regions, hydropower development has also been extensive, and several large rivers in the Canadian Shield have been dammed or their flows altered by diversion [*Déry et al.*, 2005]. The Nelson River yields 94 km^3 of flow each year, a large portion of which used to previously flow through the Churchill River to the Hudson Bay. After 1980, the hydropower projects of the Nelson were surpassed by a reservoir scheme on La Grande River in Québec, which underwent diversion and now discharges 100 $km^3\,yr^{-1}$, compared to its original natural flow of 67 $km^3\,yr^{-1}$. Other projects, like some in Norway and Northern Ontario that are less massive in scale, similarly affect river flow, with associated effects of water use on the hydrologic regime, water temperature, and ecology.

Mountains exert strong influence on atmospheric circulation and precipitation, both over synoptic and local scales. Many mountain streams are intensely modified by hydropower regulation, which in turn influences downstream ecosystems and seasonal water balances. Other effects on resources include a changing snow cover duration, which will influence recreational activities such as skiing, and also avalanches that pose hazards to transportation and settlements. Furthermore, heavy rainfall can cause landslides that affect traffic and infrastructure in the Norwegian and Cordilleran mountains.

Similar to shield and boreal regions, river systems in grasslands have in many places been subject to extensive modification and diversions [*Yang et al.*, 2004a, 2004b; *Schindler and Donahue*, 2006; *Stuefer et al.*, 2011], which influence both water exchanges with the atmosphere, final transport to the ocean, and local ecosystems and resources. In Canadian grasslands, paleoclimatic evidence suggests that the relatively moist conditions that prevailed during the twentieth century are unusual [*Sauchyn et al.*, 2002] and that the now heavily modified water systems may become subject to severe water shortages in summer if the prolonged droughts of preindustrial times were to return in combination with elevated pressures from climate change [*Schindler and Donahue*, 2006].

The mass loss of Arctic glaciers and the GrIS are presently the largest contributors to ocean storage and global sea level rise. In addition, changes in the GrIS outflow have potential ramifications for ocean thermohaline circulation around Greenland [*Rahmstorf et al*., 2005, 2015; see also *Carmack et al*., 2016]. Glaciers are important suppliers of freshwater to mountainous basins, and their shrinkage will require changed planning of hydropower operations in the long run [see *Instanes et al*., 2016]. Locally, long-term streamflow and flooding increases may be amplified in glacierized basins [*Dahlke et al*., 2012]. Spreading and thickening of aufeis [*Kane*, 1981] may be intensified due to increase of underground water storage and release, in turn caused by permafrost thaw.

Arctic wetland systems may act as accumulators of water, where it is increasingly subject to evapotranspiration. They also constitute important water storage reservoirs that regionally are strong controls on runoff formation. Most Arctic wetlands are generally species-rich ecosystems, with marshes and deltas as particularly boundary-crossing regions between terrestrial hydrological regimes and oceans. In this regard, the four largest Arctic rivers have markedly different interfaces with the Arctic Ocean. On the Lena and the Mackenzie, the river flows into the ocean through very large deltas, dominated by the river water. In contrast, the Yenisey and Ob Rivers enter the ocean through marine-dominated estuaries, where tidal flows and surges exert strong influence on local environments.

In estuaries, which are elongated and open for tides, the hydraulic regime is more dynamic than in the deltas. Sedimentation in estuaries occurs under salt-fresh water contact and biogeochemical transformation [*Gordeev et al*., 1999; *Gordeev*, 2006; *Fedorova et al*., 2015]. The marine saltwater intrusion rises higher along the river in estuaries than in deltas and also consists of a more distinct salt-fresh water geochemical barrier. The freshwater plume extending into the ocean from river mouths can therefore be said to be reversed during times in estuaries. In deltas, however, the accumulation of material can be 8–10 times higher than in estuaries. For example, the turbidity in the top of the Lena River delta is 40 mg L^{-1} on average but falls to 3–5 mg L^{-1} at the coastal line [*Fedorova et al*., 2015].

6. Major Knowledge Gaps and Future Research Directions

6.1. Processes

Arctic terrestrial precipitation trends are inherently difficult to detect given snowfall measurement challenges resulting from gauge undercatch of solid precipitation, sparsely distributed observations, low precipitation amounts, and rare long-term records [*Serreze and Hurst*, 2000; *Adam and Lettenmaier*, 2003; *Yang et al*., 2005]. Furthermore, the large natural variability of precipitation implies that a trend may take time to emerge. The compounding effects of elevation on precipitation in topographically complex regions of the Arctic, where the distribution of observing stations is biased toward low elevations and coastal regions, are also a factor [*Kane and Stuefer*, 2015]. As solid precipitation remains challenging to measure, in situ and remote sensing observations of snow cover on the ground are widely used as a precipitation metric instead of traditional precipitation gauge measurements [*Stuefer et al*., 2013]. Improving representation of precipitation and snow cover processes specific to the high latitudes, both by better remote sensing technology and more skillful models from a land surface perspective, remains an important research direction that would enhance our ability to understand and model Arctic freshwater [see *Lique et al*., 2016] and its role in the larger Earth system.

Evapotranspiration measurements in the terrestrial Arctic are mostly from field programs of short duration covering specific regions [*Langer et al*., 2011; *Mueller et al*., 2011; see also *Vihma et al*., 2016]. Because of these data gaps, analysis of evapotranspiration over the terrestrial Arctic often involves information from land surface models and remote sensing data, as well as global reanalysis and climate model outputs [*Rawlins et al*., 2010]. However, the progress in exploiting remote sensing techniques for measuring high-latitude precipitation and evapotranspiration is very slow and requires increased efforts from the hydrological community [*Lettenmaier et al*., 2015]. The complexity of modeling evapotranspiration arises from heterogeneity in land surface and a large set of parameters to consider, including soil properties, characteristics of vegetation, and availability of moisture. Studies that examine changes in land surface variables and fluxes [e.g., *Destouni et al*., 2013; *Dirmeyer et al*., 2013; *Jaramillo and Destouni*, 2014; *Törnqvist et al*., 2014; *van der Velde et al*., 2014; *Bring et al*., 2015a], but focus on the terrestrial Arctic, may improve our picture of the land-atmosphere interactions with respect to soil moisture and evapotranspiration. In general, there is a great need for more accurate representation of land surface effects, such as vegetation, lakes, and human interventions, in climate models [*Harding et al*., 2014; *Bring et al*., 2015a].

Knowledge of runoff variability in the Arctic is limited by river observations, which cover about 70% of the total contributing area [*Lammers et al*., 2001; *Shiklomanov et al*., 2002, 2006; *Bring and Destouni*, 2013], and whose availability and reliability are even more lacking when considering water chemistry and sediment [*Holmes et al*., 2000, 2012; *Prowse et al*., 2005; *McClelland et al*., 2008, 2015; *Bring and Destouni*, 2009]. Notwithstanding these limitations, our understanding of how discharge changes in response to land surface and permafrost changes is not complete. Observed divergence in precipitation and discharge trends across basins [e.g., *Berezovskaya et al*., 2004; *Bring and Destouni*, 2011] indicate that changes in surface and subsurface water interactions, and their feedback on water balance and evapotranspiration [*Bosson et al*., 2013], need to be better understood and quantified, although in some cases such discrepancies are due to data uncertainty. Climate model agreement on runoff changes is also particularly limited for Arctic basins [*Bring and Destouni*, 2011; *Törnqvist et al*., 2014]. Along the coast, information shortages combine to yield great uncertainty both in surface flows, which are poorly monitored for many smaller rivers, and the interaction between groundwater, surface water, permafrost, and seawater [*Bobba et al*., 2012; see also *Carmack et al*., 2016]. To address these knowledge gaps in river fluxes, some proposed strategies to improve monitoring may help [e.g., *Karlsson et al*., 2011; *Azcárate et al*., 2013; *McClelland et al*., 2015], as will a continuation of the Sustaining Arctic Observation Networks process. The potential of remote sensing methods, as a complement to ground observations, will likely increase with new instruments (e.g., the NASA Surface Water and Ocean Topography mission and the EU Sentinel missions), and some promising results so far [e.g., *Smith and Pavelsky*, 2008; *Biancamaria et al*., 2011; *Gleason and Smith*, 2014] indicate that the Arctic hydrology community should investigate these opportunities further. Another remaining knowledge gap is how to ensure efficient long-term monitoring of Arctic rivers under changing climate conditions, something that should be addressed by combining climate projections and network optimization methods. Furthermore, updates are now required to the rating curves for major Arctic gauges, without which a much desired coordinated monitoring of water temperature, chemistry, and suspended matter will be hampered [*McClelland et al*., 2015]. Hydrology-ecology interactions should be investigated through field monitoring and new methods using isotopes and remote sensing. Multidisciplinary investigations at the drainage basin scale should be encouraged as they have the potential to increase understanding of interaction between hydrology and atmosphere, oceans, ecology, and resources [see also *Wrona et al*., 2016].

Main knowledge gaps in permafrost hydrology include the understanding of underground and surface water interaction (for example, inflow and outflow from riverbed taliks), the mechanisms of infiltration and flow formation in frozen and thawing ground in different landscapes, evaporation processes, atmosphere-land surface-subsurface energy interaction, and the patterns of precipitation distribution at catchment scales (including snow redistribution and intensive rain events in mountainous conditions). Scale discrepancies between field studies and catchments, and between hydrological models of various scales and climate data, are a challenge. Runoff formation governed by active layer thaw-freeze processes varies spatially at slope scale, and modeling and parameter estimation, even for large basins, require consideration of this. Advances in modeling these processes, ideally in collaboration with the soil carbon community, would be a key step toward understanding how wetting and drying of permafrost landscapes will influence soil carbon release to the atmosphere during the present century [*Schuur et al*., 2015]. Observational data uncertainty, especially for extreme conditions, remains an obstacle to advances in understanding. However, existing catchment study data are sometimes underutilized. Additional parameters for small and large models can be assessed based on field measurements and improved equipment. To capitalize on these underexploited opportunities, integrating data and studies across the Arctic should be a priority.

6.2. Hydrophysiographic Regions

An overarching and pressing research question across all regions of the Arctic is how to separate anthropogenic from natural disturbances. Human impacts are not easily quantified as the type and intensity of activities are difficult to predict, and their effects on hydrology are to a degree uncertain, particularly on a larger systems level. The potential for new methods that distinguish change drivers, previously applied globally to separate landscape and atmospheric components [*Jaramillo and Destouni*, 2014], should be investigated also for the Arctic region.

Anthropogenic activities are particularly likely to increase in the coastal tundra zone [*Andreeva*, 1998]. However, the land-ocean boundary commonly falls in between central monitoring and observation strategies that prioritize information goals based on either a terrestrial or a marine perspective [*Strandmark et al*., 2015]. An explicit focus on the hydrological interactions between land and sea across all of the coastal interface, and not just through rivers, would help overbridge disciplinary boundaries and expand knowledge on a relatively understudied part of the Arctic freshwater system. Also, our understanding of marsh hydrology, including hydrodynamics and hydroecology of the Arctic coastal zone, is limited [*Overduin et al.*, 2014], as well as our knowledge of submarine groundwater discharge and the forming of salt-fresh aquifers in coastal zones [*Bobba et al.*, 2012]. Groundwater models of varying complexity [e.g., *Danielescu et al.*, 2009; *Mazi et al.*, 2013], potentially coupled to hydrological and oceanographic models, should be applied to Arctic conditions to increase understanding of the coastal zone.

Large parts of the boreal zone are underlain by discontinuous permafrost, and these areas are also prone to development in the near future. The anthropogenic impact on the hydrological cycle from forest clearing and fires is not fully understood. More information is needed on the present status of groundwater circulation and storage in discontinuous permafrost terrain in order to understand impacts on heat and water fluxes and reasonably project future conditions. Emerging and improved remote sensing techniques will enhance the opportunity for surveillance of surface wetness on a regional scale [e.g., *Leconte et al.*, 2008]. Future studies of mountain hydrology are also needed to determine if those areas will continue to provide as large a share of river flow to the Arctic Ocean as today.

Projections of future precipitation changes constitute a key hydrological uncertainty in grasslands [*Zhang et al.*, 2011]. In addition, the previously noted uncertainty in future drought patterns [*Schindler and Donahue*, 2006] will strongly influence grassland hydrology. To improve knowledge and improve adaptation efforts, improved downscaling of drought-related variables, as well as further exploration of teleconnection patterns that govern droughts [*Schindler and Donahue*, 2006], is necessary. In grasslands, the extensive human modifications of the water cycle should also be considered in both model projections of future changes [*Bring et al.*, 2015a] and in the evaluation of past changes [*Jaramillo and Destouni*, 2015].

In terms of glaciers and ice sheets, dominant unknowns and uncertainties in modeling and understanding of ice sheet mass balance and runoff response are percolation, retention, and freezing of melt and rain water by snow [e.g., *Box*, 2013; *Vernon et al.*, 2013]. Extensive surface melt has been recently documented over the GrIS, unprecedented in the observational record (NOAA Arctic report card 2012 available at www.arctic.noaa.gov/reportcard) [*Nghiem et al.*, 2012; *Hanna et al.*, 2014]. Meltwater can be trapped within the firn, and so the magnitude and delay of runoff remain unknown. Conventional thought has been that practically all air space in the snow must be filled before meltwater can exit the ice sheets as runoff. Yet recent snow and ice core measurements provide the first evidence of the formation of impermeable ice layers, enabling an abrupt increase in ice sheet runoff [*Vernon et al.*, 2013]. Other recent results indicate that ice sheet outflow may be overestimated in climate models due to omission of subglacial processes [*Smith et al.*, 2015]. Development of percolation/retention schemes and evaluation of nonlinearities in ice sheet meltwater runoff should therefore lead to more accurate projections of GrIS runoff.

To increase our understanding of wetland dynamics, there is a need for improved riverbed morphological parametrization for hydraulic models, especially for large Arctic rivers. Such data would also improve the ability to model river fluxes, in general, and not just wetlands. In particular, there is a need to monitor the seasonal and interannual changes in wetlands, as near-surface moisture conditions strongly affect evaporation and can influence the development of precipitation and even summer weather. Evapotranspiration measurements are critical to understand water fluxes, and organization of such measurements across different wetland and peatland types is required. There is also a need for method improvements, including models, remote sensing, and isotopic techniques [*Tetzlaff et al.*, 2015].

7. Conclusion

Terrestrial freshwater pathways have a central role in water, material (ions), and energy exchanges between components of the Arctic freshwater system. They connect the atmosphere and ocean over large distances, along which ecosystems and resources are affected by varying geographical characteristics of the terrestrial freshwater systems.

Across the terrestrial Arctic, snow cover is generally decreasing, but patterns are complex in terms of duration and extent, and the snow water equivalent. Evapotranspiration is coupled to landscape and permafrost changes in different ways across the Arctic, with wetting and drying occurring in succession, depending on the permafrost conditions. However, some regions may undergo systematic shifts in one direction or the other. Annual river flux patterns show a long-term increase at the pan-Arctic scale, and patterns of higher discharge also seem to apply more universally across regions in recent decades. Reports of permafrost warming are numerous, and an expected effect of increasing river base flows has been reported for several regions. River and lake ice cover have decreased in duration, with the most rapid reductions observed from the late twentieth century.

Water flux and storage change in complex ways across Arctic regions. Permafrost degradation will occur both in the tundra and boreal zone, with complete permafrost degradation occurring along the margins of the permafrost zone in the taiga. Evapotranspiration and precipitation are likely to increase in all regions, although summer precipitation may decrease in the southern parts of the Arctic. River flows will likely increase on average across all regions, but high and low flows may decrease in parts of the Arctic boreal plains, shields, and grasslands regions, mostly where local decreases in average flows are also projected.

To improve our understanding of changes to the freshwater system in the Arctic, a number of knowledge gaps are of particular importance. We are still lacking a comprehensive picture of evapotranspiration across Arctic regions and how it will interact with climate and landscape changes. These changes are coupled to precipitation, river flows, and groundwater flows, which underlines that it remains a challenge to estimate full water budgets for most Arctic river basins. For coastal regions, the interaction of flow along various poorly constrained pathways implies that these regions are essentially uncharted. As some parts of the Arctic increasingly become subject to settlement and natural resource extraction, it will become more important to separate anthropogenic effects from underlying environmental changes.

To address these challenges, future research should include more widespread use of multimodel ensembles to enable better separation of model-internal variability from change. Novel approaches to overcome data limitations will be needed, as they are not likely to diminish appreciably without large-scale funding increases. Concerted efforts at studying water, solute, and energy fluxes on various scales will remain important to detect, understand, and project water system changes. Despite a growing potential of remote sensing, it remains critical to at least maintain the current ground-based capacity to observe Arctic terrestrial hydrology.

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