



Evaporation Processes and Changes Over the Northern Regions

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Yinsheng Zhang, Ning Ma, Hotaek Park, John E. Walsh, and Ke Zhang

Abstract

Evapotranspiration (ET) is a key component in global water and energy cycles. This chapter presents and discusses recent research advances about ET over northern regions and watersheds. ET in northern regions tends to increase with the decrease of latitude. The largest ET typically appears in the forest ecosystem, while the grasslands and shrublands have small ET. While the seasonal variations in ET are usually high, the interannual variability in annual ET is usually low over the Arctic regions. Sublimation from snow cover accounts for about 15–25% of winter precipitation. Many factors, such as soil moisture, vegetation type and productivity, and ecosystem features affect ET over northern regions. In addition, precipitation plays a key role in impacting ET. ET is more sensitive to precipitation in the early growing season than in the late growing season. Furthermore, changes in freeze–thaw processes due to warming also affect land surface conditions and the ET processes. During 1983–2005, ET

Y. Zhang (✉) · N. Ma

Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing, China
e-mail: yinshengzhang@icloud.com; yszhang@itpcas.ac.cn

N. Ma

e-mail: ningma@itpcas.ac.cn

H. Park

Japan Agency for Marine–Earth Science and Technology (JAMSTEC), Yokohama, Japan
e-mail: park@jamstec.go.jp

J. E. Walsh

International Arctic Research Center, University of Alaska, Fairbanks, AK, USA
e-mail: jewalsh@alaska.edu

K. Zhang

College of Hydrology and Water Resources, Hohai University, Nanjing, China
e-mail: kzhang@hhu.edu.cn

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increased significantly in the Arctic region with a rate of 3.8 mm decade⁻¹ because of regional warming and vegetation greening. Such an increase in ET may exert significant impacts on the regional hydrology and water resources. Advanced models can simulate past ET change over the large northern watersheds. Remote sensing has provided new ET data and information that support climate and hydrology research and applications. There is a key question: Will Arctic landscapes become wetter or drier as climate changes? According to global models and data analyses, annual ET has increased over the northern regions. In the future, summer PE is projected to decrease much of Canada, increase over Alaska, decrease over the western and northern Eurasian subarctic, and increase over parts of northeastern Russia. Over most of these areas, the sign of the projected change is not robust across the models at the 95% confidence level. Many factors contribute to the uncertainty in the projected changes in Arctic surface wetness. There is certainly a need to better quantify and narrow the uncertainties in global models in the northern regions.

4.1 Introduction

Evapotranspiration (ET) refers to the amount of water vapor evaporated from the unit area of the land surface during a unit of time and consists of evaporation from canopy-intercepted water, evaporation from soil (E), and transpiration from plants (T). ET is a key component in global water and energy cycles. On average, more than 60% of global land precipitation is returned to the atmosphere through terrestrial ET (Oki and Kanae 2006). ET as latent heat flux consumes roughly 50% of the solar radiation absorbed by the Earth's surface (Trenberth et al. 2009). ET affects climate through a wide range of feedbacks to air temperature, humidity, and precipitation (Shen et al. 2015; Shukla and Mintz 1982; Zeng et al. 2017). Accurate estimation of ET is therefore fundamental to not only elucidating how the hydrological cycle responds to climate change but also to regional drought monitoring and water resources management (Fisher et al. 2017; Ma et al. 2019). However, investigating ET is challenging because of its complex interactions across the soil–vegetation–atmosphere interface (Katul et al. 2012; Matheny et al. 2014; Zhang et al. 2014). Historically, ET was related (via linear or nonlinear scaling with soil moisture) to direct measurements of its conceived maximum value, i.e., pan evaporation (Brutsaert 2013). However, neither water-limited nor energy-limited land surface ET acts as pan evaporation (Brutsaert 1982). With recent advances in instrumentations and data storage, a better understanding of ET from various land covers has been reported through the use of in situ flux observations (e.g., Baldocchi et al. 2004; Fischer et al. 2013; Ma et al. 2014; Mackay et al. 2007; Wilson and Baldocchi 2000). These studies have significantly contributed to our knowledge of regional hydrological regimes and their complex feedback mechanisms between land and atmosphere (Baldocchi 2014).

Because transpiration from plants (T) links the water and carbon cycles, it is used to calculate carbon assimilation by terrestrial vegetation, so estimating T fluxes is a major focus in climate and ecology studies (Evaristo et al. 2015; Jasechko et al. 2013; Kool et al. 2014; Wei et al. 2017). As T process directly correlates with plant growth and the carbon cycle (Scott et al. 2006), quantitative estimation of T in the total evapotranspiration (T/ET) has long been acknowledged to play a crucial role in water resource management, yield estimation, water cycle, and climate change, from plot scale to global scale (Schlesinger and Jasechko 2014; Scott and Biederman 2017; Xiao et al. 2018). There is, however, a considerable discrepancy among global T/ET estimations by different methods. For example, the results of a combination of wide ranging, remotely sensed observations showed that approximately 80% of the annual land ET is attributed to T (Miralles et al. 2011). Wei et al. (2017) quantified the global T/ET with a leaf area index (LAI)-based ET partitioning algorithm and concluded that T accounts for 57% of ET. The distinct isotope effects of T and E based on the isotopic analysis of a global dataset of large lakes and rivers showed that T represents 80–90% of terrestrial ET (Jasechko et al. 2013), although this estimate was challenged by Coendersgerrits et al. (2014). The results of isotope mass budget-based simulations suggested that the transpired fraction of ET accounts for approximately 60% of the annual land ET (Good et al. 2015). There are also great challenges in the state-of-the-art land surface models and remote sensing models in representing the ratio of T to ET. Maxwell and Condon (2016) argued that partitioning ET is connected to water table depth and they found that including lateral groundwater flow in the model increases transpiration partitioning from 47 ± 13 to $62 \pm 12\%$. This aspect was recently also acknowledged in Chang et al. (2018), in which terrain-driven lateral water flows spread out soil moisture to a wider range along hill slopes with an optimum sub-range from the middle to upper slopes, where soil evaporation was more suppressed by the drier surface than T due to plant uptake of deep soil water, thereby enhancing T/ET. In terms of remote sensing models, large errors in representing the components of ET also exist in Moderate Resolution Imaging Spectroradiometer (PM-MODIS), the Priestley–Taylor Jet Propulsion Laboratory model (PT-JPL), and the Global Land Evaporation Amsterdam Model (GLEAM), which shows root-mean-square-error of 90–114% for soil evaporation and 54–114% for transpiration (Talsma et al. 2018).

This chapter reviews recent research and results of ET from various land surfaces (including different vegetation types) and also across large regions and watersheds. Specifically, it discusses ET process, pattern, and variabilities over space and time, such as regional/basin ET, its change, and impact to water balance. It also demonstrates model estimates/simulation of large-scale ET over the arctic domain and selected watersheds, highlights remote sensing development in ET estimation, and global model analysis of the net moisture flux (P-E) and its change across the northern regions.

4.2 ET Distribution and Variability

Atmospheric conditions, such as wind speed and saturation deficit, are dominant factors in determining ET variability. Zhang et al. (2003) summarized that daily mean evaporation over the Tibetan Plateau varied within the range of 0.3–3.5 mm on the permafrost surface, and regional differences in evaporation were strongly related to surface soil moisture. Locally, topography and its influence on surface soil moisture was found to control evaporation systematically. The seasonality of evaporation in permafrost regions is dominated by freeze–thaw cycles at the surface; evaporation from the melting permafrost surface is up to 4–7 times greater than that from the frozen ground. In forested terrain, the interception of precipitation can reduce daily evaporation by 60–70%. Sublimation from the snow surface observed at Tianshan Mountains and eastern Tibetan Plateau was in the range of 0.2–1.0 mm d⁻¹. Mu et al. (2009) investigated the spatial pattern of evapotranspiration for the pan-Arctic domain and found large annual ET variability during 2000–2006 among the regional biomes. The largest annual ET rates occur over forests, while the lowest rates occur over grasslands and scrublands; annual ET rates for savanna and cropland areas are generally intermediate. Ecosystem processes in high-latitude boreal and tundra biomes are strongly constrained by low solar irradiance and freezing temperatures for much of the year, so that seasonal patterns in plant photosynthesis (GPP) and ET correspond closely and are generally confined to a relatively narrow growing season. The interannual variability in estimated annual ET is relatively low although the seasonal variation is high over the Arctic, which reflects the dominance of cold temperature constraints on boreal Arctic ecosystem processes (Fig. 4.1).

Variation in evapotranspiration is influenced by all ecosystem parameters and processes, such as soil moisture content, vegetation productivity, and ecosystem nutrient and water budgets. The partitioning of available energy into evapotranspiration (latent heat flux) and sensible heat flux at a vegetation surface also affects aspects of weather and climate. As ET consists of soil evaporation, and transpiration from plants, the soil moisture, and biological feature are key factors to determine ET variation. Zhang et al (2005) demonstrated that E is more sensitive to surface soil moisture than T. The former takes water from a very thin soil layer and moves it to the atmosphere; the latter transfers water from the soil through root–stalk–leaf mechanisms and may take water from a thicker soil layer. Therefore, T is anticipated to be more independent of soil moisture variations than E. Figure 4.2 shows the relationship between E/Ep (Ep is potential evaporation) and the surface soil moisture. E/Ep increased almost linearly with ground surface moisture when the volumetric water content was less than 30% but varied little when moisture content beyond that level. This suggests that soil evaporation was constrained by the deficiency of available water when soil moisture was less than the critical value of 30%, illustrating the primary controls of soil water status on the soil evaporation.

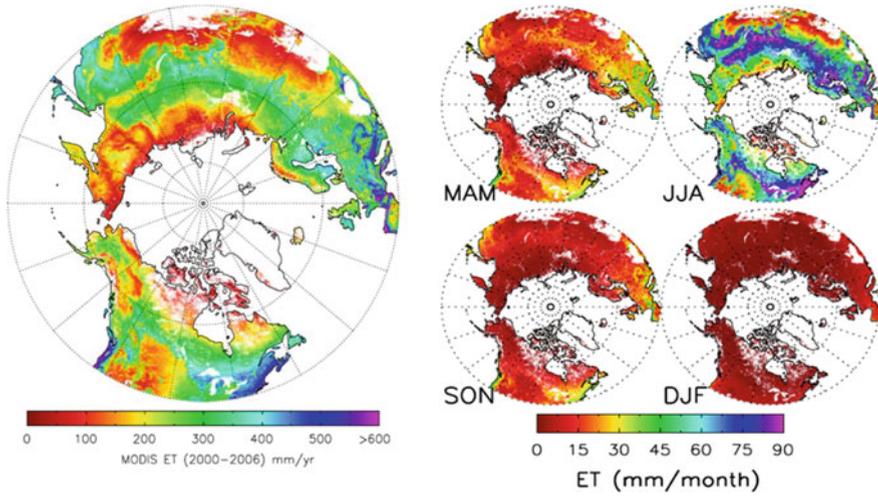


Fig. 4.1 Spatial pattern of mean annual and seasonal ET during 2000–2006 for the pan-Arctic domain (Mu et al. 2009)

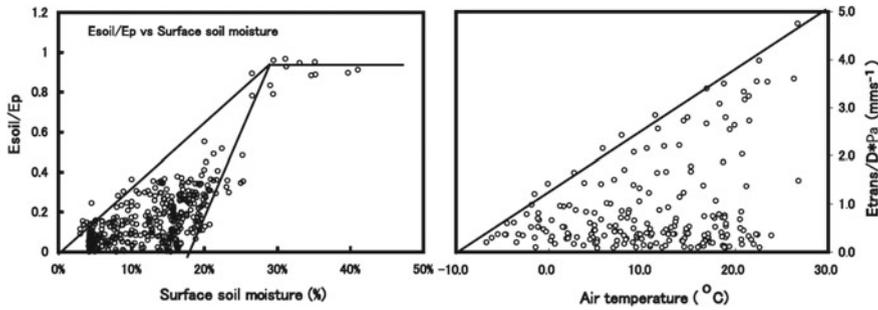


Fig. 4.2 Relationships between (left) E/E_p and surface moisture and (right) leaf conductance ($E_{trans}/D * Pa$, mms^{-2}) and air temperature. D , leaf-to-air specific deficit; Pa , air density (After Zhang et al. (2005) for sparse grassland in northeastern Mongolia)

Transpiration rates are determined by two parameters: leaf conductance and leaf-to-air specific deficit (D). While it is impossible to present leaf conductance because of no CO_2 concentration data were collected, the right panel of Fig. 4.2 shows the ratio of $E_{trans}/D * Pa$ (D is the leaf-to-air specific deficit and Pa is air density), which implies a leaf conductance value, plotted against air temperature. The plots show that transpiration increased as temperatures increased. When temperatures became warm, higher leaf conductance allowed water uptake if soil moisture was low but higher than wilting point (2.8–4.1% in the root zone). Lower temperatures, lower D , and higher humidity accompany precipitation events, so

plant transpiration decreased as soil evaporation approached to potential evapotranspiration (Fig. 4.2).

Projected increases in air temperature and precipitation due to climate change in Arctic wetlands could dramatically affect ecosystem function. As a consequence, it is important to define controls on evapotranspiration, i.e., the major pathway of water loss from the system. Liljedahl et al. (2011) quantified the multi-year controls on midday Arctic coastal wetland evapotranspiration, measured with the eddy covariance method at two vegetated, drained thaw lake basins near Barrow, Alaska. Variations in near-surface soil moisture and atmospheric vapor pressure deficits were found to have nonlinear effects on midday evapotranspiration rates. Data collections near Barrow over 3 years showed that vapor pressure deficits (VPD) near 0.3 kPa appeared to be an important hydrological threshold, allowing latent heat flux to persistently exceed sensible heat flux (Fig. 4.3). Dry (compared to wet) soils increased bulk surface resistance (water-limited). Wet soils favored ground heat flux and therefore limited the energy available to sensible and latent heat flux (energy-limited). Thus, midday evapotranspiration was suppressed from both dry and wet soils but through different mechanisms. They also found that wet soils (ponding excluded) combined with large VPD, resulted in an increased bulk surface resistance and therefore suppressing evapotranspiration below its potential rate (Priestley-Taylor coefficient < 1.26). This was likely caused by the limited ability of mosses to transfer moisture during large atmospheric demands. Ultimately, in addition to net radiation, the various controlling factors on midday evapotranspiration (i.e., near-surface soil moisture, atmospheric vapor pressure, and the limited ability of saturated mosses to transfer water during high VPD) resulted in an average evapotranspiration rate of up to 75% of the potential evapotranspiration rate. These multiple limitations on midday evapotranspiration rates have the potential to moderate interannual variation of total evapotranspiration and reduce excessive water loss in a warmer climate. Combined with the prevailing maritime winds and projected increases in precipitation, these curbing mechanisms will likely prevent extensive future soil drying and hence maintain the presence of coastal wetlands.

Evapotranspiration dominates hydrological processes on the Arctic Coastal Plain for a couple of months after snowmelt until soil moisture declines (Kane et al. 2000, 2008). The majority of studies on ET in the Arctic tundra focus on whole-ecosystem fluxes, with rates of approximately 1–3 mm day⁻¹ (Liljedahl et al. 2011; Mendez et al. 1998). However, a whole-ecosystem approach to determining ET rates does not allow for quantifying the variability in fluxes associated with the heterogeneous landscape (Oren et al. 2006), particularly on the Arctic Coastal Plain (Oechel et al. 1998). Further, a whole-ecosystem approach does not allow for partitioning ET into its components of evaporation and transpiration. Spatial heterogeneity in soil moisture, soil temperature, and plant composition likely affect how ET is partitioned into evaporation and transpiration in the Arctic Coastal Plain (Oberbauer and Dawson 1992). It is critical to understand the partitioning of evapotranspiration because environmental processes control evaporation and transpiration differently (Jasechko et al. 2013). While both respond to surface energy, atmospheric demand, and soil water availability (Betts et al. 1999;

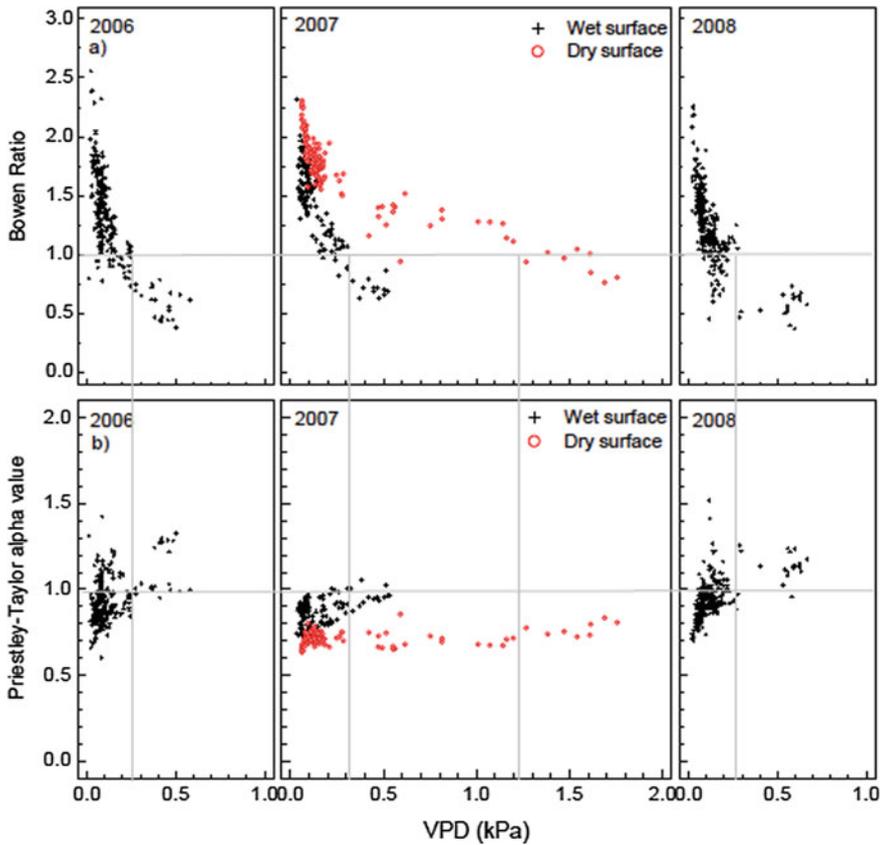


Fig. 4.3 The relationship between mean hourly air vapor pressure deficit (VPD) and **a** Bowen ratio (β) or **b** Priestley–Taylor α during differing soil moisture conditions at the BE site, 2006–2008. Dry soils represent a soil water potential < -0.13 MPa at 10 cm depth. The vertical dashed lines represent the identified critical value of VPD. VPDs above this threshold resulted in a < 1 and a Priestley–Taylor α near or above 1. The identified VPD thresholds were 0.25 (2006), 0.31 (2007), and 0.28 kPa (2008) for wet soils and 1.19 kPa for dry soils (2007). (© Liljedahl et al. 2011, distributed under the Creative Commons Attribution 3.0 License. use with permission from the authors)

Calder 1998), evaporation is a physical process but transpiration is a plant physiological process controlled by stomata (Wullschlegel et al. 1998).

On the tundra surface near Barrow, Alaska, Raz-Yaseef et al. (2017) found that variation in environmental conditions and plant community composition, driven by micro-topographical features, has a significant influence on ET. ET had high variability across the field site, i.e., the fluxes were highest over mosses and open water, lower from grasses and sedges (65% of those from mosses and open water), and lowest over bare ground and lichens (50% of those from mosses and open water) (Fig. 4.4).

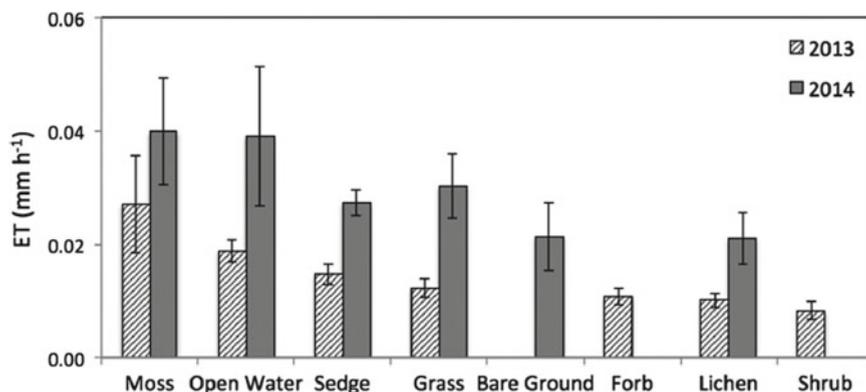


Fig. 4.4 Two years ET measured with the portable chamber. Values are averages for each plant type and for open water; bars denote standard deviation (After Raz-Yaseef et al. 2017)

Among plant types, ET from moss and inundated areas was more than twice that from other plant types. ET from troughs and low polygonal centers was significantly higher than that from high polygonal centers. ET also varied seasonally, with peak fluxes of 0.14 mm h^{-1} in July. Diurnal fluctuations in incoming solar radiation and plant processes produced a diurnal cycle in ET. Observed patterns with projections for the impact of permafrost degradation on polygonal structure suggest that micro-topographic changes associated with permafrost thaw have the potential to alter the tundra ecosystem ET (Young-Robertson et al. 2018). Yuan et al. (2010) examined the impacts of precipitation seasonality and ecosystem types on ET quantified by eddy covariance towers from 2002 to 2004 in three ecosystems (grassland, deciduous broadleaf forest, and evergreen needle leaf forest) in the Yukon River Basin, Alaska. The annual precipitation changed greatly in both magnitude and seasonal distribution through the three investigated years. Observations and model results showed that ET was more sensitive to precipitation scarcity in the early growing season than in the late growing season, which was the direct result of different responses of ET components to precipitation in different seasons. The results demonstrated the importance of seasonal variations of precipitation in regulating annual ET and overshadowing the function of annual precipitation. Comparison of ET among ecosystems over the growing season indicated that ET was largest in deciduous broadleaf, intermediate in evergreen needle leaf, and lowest in the grassland ecosystem. These ecosystem differences in ET were related to differences in successional stages and physiological responses.

Sublimation from snow surface has been identified as an important hydrological process at high altitudes and in high-latitude regions, involving complex mass and energy exchanges. In the Colorado Frontal Ranges, measurements from snow evaporation pans indicated that net sublimation for the 5-month winter period from December to April was 135 mm (Meiman and Grant 1974). Berg (1986) estimated

sublimation losses from snow cover to be 30–51% of precipitation over a 2-year period from 1973 to 1975. Kattelmann and Elder (1991) estimated sublimation from snow to be 18% of total precipitation over 2 years in the Sierra Nevada. In the subarctic region, several studies have shown that sublimation from snow cover is a non-negligible hydrological component that also affects river discharge and regional water resources. In western Canada, sublimation from snow during the winter season consumed from 15 to 40% of seasonal snowfall (Woo et al. 2000) and 12 to 33% of annual snowfall (Pomeroy and Li 1997). Suzuki et al. (2002) estimated that sublimation from snow cover in eastern Siberia was significant at 25.6% of precipitation from October to April. However, these estimations were derived from modeling and lack of observational verification. Subarctic ground surfaces are dominantly covered by subalpine and boreal forests. The impact of forests on snow cover has been extensively investigated via accumulation and melting processes. An increase of 30–45% in seasonal snow accumulation was measured after the removal of the evergreen forest (Pomeroy and Gray 1995; Pomeroy and Li 1997). Pomeroy et al. (1998) also found that snow water equivalent (SWE) generally increases with evergreen canopy density in boreal forests. Simulation models for snowmelt under a forest canopy have been developed to examine the relationship between snowmelt and forest density (Barry et al. 1990; Yamazaki and Kondo 1992; Wigmosta et al. 1994). Canopy density is important in controlling snow ablation timing and rates because tree height and canopy properties control the transmission of solar radiation (Davis et al. 1997; Ni et al. 1997).

Leydecker and Melack (1999) discussed the sensitivity of sublimation over a short time scale using snow surface roughness, instrument height, and wind speed and demonstrated that wind speed was the critical variable for determining sublimation and that doubling wind speed can triple the sublimation. Zhang et al. (2008) demonstrated that when the wind speed is less than 2.0 m s^{-1} , the saturation deficiency is predominant in determining sublimation, and sublimation increases significantly when the wind speed is above 2.0 m s^{-1} . The variation in vapor transfer coefficient versus wind speed can be deduced (Fig. 4.5). This result shows that the vapor transfer coefficient for the snow surface is larger under low wind and decreases sharply as wind speed increases. However, the vapor transfer coefficient is nearly constant when the monthly mean wind speed is above 2.0 m s^{-1} . The impact of the coefficient on the sublimation rate is not so clear. Daily sublimation shows a tendency toward correlation with the vapor transfer coefficient (Fig. 4.5). The values of 20.3–21.6% of total snowfall lost to sublimation compares favorably with snow cover lost due to sublimation of 25.6% in eastern Siberia (Suzuki et al. 2002), or 12–33% of annual snowfall in Canada (Pomeroy and Li 1997).

Snow cover models were used to quantify sublimation. Sexstone et al. (2018) demonstrated that snow sublimation rates corresponding to climate-warming simulations remained unchanged or slightly increased but total sublimation losses decreased by up to 6% because of a reduction in snow-covered area and duration. Seasonally snow-covered forests in western North America have experienced substantial disturbance from mountain pine beetle (*Dendroctonus ponderosae*) and

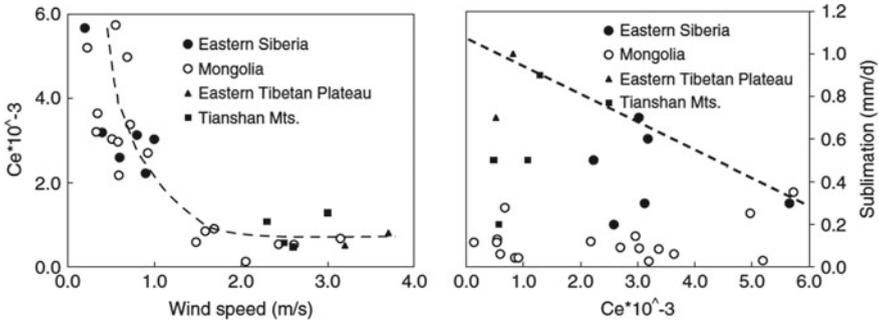


Fig. 4.5 Vapor transfer coefficient versus wind speed (left) and daily sublimation versus the vapor transfer coefficient (right) (Zhang et al. 2008)

spruce beetle (*Dendroctonus rufipennis*) outbreaks (Potter and Conkling 2016), which have resulted in widespread tree mortality and thus changes to forest structure that are particularly relevant to canopy and surface sublimation processes. Field studies that have focused on measuring SWE in both unimpacted and disturbed forests have inferred decreasing (e.g., Boon 2012; Pugh and Small 2012) as well as steady or increasing (e.g., Biederman et al. 2014; Harpold et al. 2014) net sublimation fluxes in the presence of disturbance. In contrast, distributed watershed modeling studies considering beetle-induced forest mortality (Penn et al. 2016) have generally reported decreased evapotranspiration but have not specifically focused on the sublimation component of evapotranspiration. Although changes to snow accumulation and melt processes from climate warming have been studied widely (Musselman et al. 2017a, b), the response of sublimation to climate change has received little investigation. As a result, specific knowledge gaps remain including how the components of sublimation will individually and collectively respond to changes in forcing mechanisms, in addition to changing snow accumulation and melt dynamics. Process-based snow models that can integrate responses and feedbacks of the snow energy balance offer the ability to evaluate sublimation responses to these changing land cover and climate conditions, which is critically important for the understanding of the water balance in snow-dominated regions.

4.3 Climate Warming and ET Change

The terrestrial cryosphere covers approximately 66 million km² (~52.5%) of the global land area where water is either permanently or seasonally frozen. Each year, most of the terrestrial cryosphere undergoes a seasonal transition between predominantly frozen and non-frozen (i.e., thawed) landscape conditions; these transitional events are relatively abrupt and analogous to a biospheric and hydrological

on/off switch between active ecosystem processes during the growing season and largely dormant conditions during the seasonal frozen period. The cold temperatures and seasonal frozen conditions pose strong constraints to vegetation productivity; surface hydrological processes, surface energy fluxes, and land-atmosphere trace gas exchange in these regions. In seasonally frozen environments, vegetation photosynthetic activity and associated ET are constrained by low temperatures and chemical unavailability of water as a result of being frozen.

Recent warming has altered freeze–thaw (FT) processes in the northern regions, resulting in substantial changes in terrestrial ecological and hydrological processes, including generally earlier onset and lengthening of potential vegetation growing seasons, thawing and degradation of permafrost, earlier spring snowmelt and increased groundwater to stream discharge from permafrost thawing. Warming and associated lengthening of the potential growing season has led to positive impacts on ecosystems of more cold-limited regions of the northern high latitudes, including increases in vegetation structure, greenness, productivity, and net CO₂ exchange. Recent warming trends also have been associated with negative ecosystem impacts within more water-limited regions of the northern latitudes by increasing water stress, which has been attributed to be a major cause of recent vegetation browning and productivity reductions, higher tree mortality rates, and increasing fire frequency.

Seasonal transitions of the landscape between predominantly frozen and thawed conditions are analogous to a biospheric and hydrological on/off switch, with marked differences in ET, vegetation productivity, and other biological activity between largely dormant winter and active summer conditions. By analyzing independent FT and ET records derived from satellite remote sensing, Zhang et al. (2011) investigated changes in FT seasons and ET from 1983 to 2006 and their connections in the northern cryosphere. They found that annual ET shows predominantly negative correlations with the timing of primary seasonal thaw (T_{thaw}) but has positive correlations with the timing of primary seasonal freeze (T_{freeze}) in most areas of the domain (Fig. 4.6). The correlations between annual ET and T_{thaw} are generally higher than that between ET and T_{freeze} , indicating that the onset of the non-frozen season has a relatively larger impact on annual ET than its cessation. This is likely because incident solar radiation available for evaporation is generally higher during the spring T_{thaw} period than the autumn T_{freeze} period; the T_{thaw} period also coincides with the general onset of the active growing season in northern, boreal, and Arctic environments, whereas vegetation photosynthetic activity and transpiration is reduced in autumn when plants are transitioning to winter dormancy (Kimball et al. 2004). Annual ET is positively correlated with the annual number of non-frozen days (DNF) over most of the domain, with stronger correlations at higher latitudes than in lower latitude areas.

In contrast to the prevailing pattern, some regions in the southern portion of the Arctic region show positive correlations between annual ET and T_{thaw} (Fig. 4.6a), where earlier than the normal onset of non-frozen period corresponds with less-than-normal annual ET. Almost all of these regions are located in areas where ET is predominantly limited by water supply. This suggests that an advancing

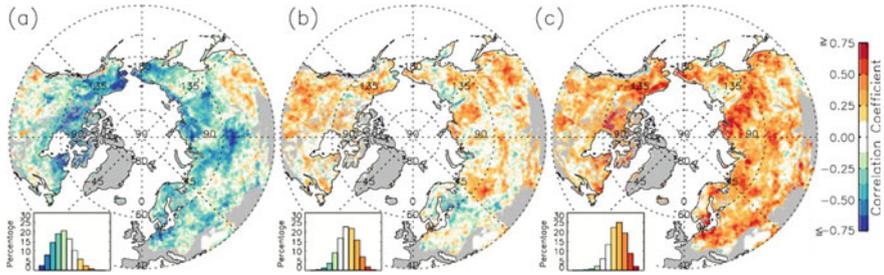


Fig. 4.6 Correlation maps **a** between annual ET and Tthaw, **b** between annual ET and Tfreeze and **c** between annual ET and DNF from 1983 to 2006; the insets show the frequency distributions of corresponding correlation coefficients, (Zhang et al. 2011)

non-frozen period may not promote additional increases in annual ET in these water supply-limited regions.

At northern high latitudes, cold temperatures and limited photoperiod largely constrain the potential growing season to a relatively brief period during the spring and summer months. Vegetation is largely dormant during the winter frozen period and coupled with frozen temperatures, seasonally low solar radiation loads, and limited day length strongly constrain ET. Low temperatures also inhibit plant photosynthesis and respiration by decreasing enzyme activity and protein synthesis in plant cells (Raich and Schlesinger 1992). Because transpiration is a consequence of plant photosynthesis, the constraint of low temperatures on photosynthesis also functions as a constraint on ET. Low temperatures also restrict photosynthesis and canopy gas exchange by limiting water supply and mobility in roots and xylem, leading to canopy stomatal closure (Woodward and Kelly 1997). Moreover, the atmospheric capacity to hold moisture increases exponentially with increasing temperature according to the well-known Clausius–Clapeyron relationship. Low temperatures, therefore, suppress ET by reducing atmospheric moisture demand. In addition, vegetation in cold boreal and Arctic regions are well adapted to the characteristic harsh environmental conditions and recover photosynthesis and respiration at relatively high rates following snowmelt and the new release of water in the landscape (Havranek and Tranquillini 1995; Suni et al. 2003; Kimball et al. 2004). Earlier onset of springtime thaw facilitates earlier photosynthesis and transpiration and generally enhances plant growth and transpiration when plant-available moisture is not limiting. During later stages of the non-frozen season in autumn, plants begin to enter the stage of senescence and dormancy, coinciding with seasonal reductions in temperatures, solar radiation, and photoperiod; thus, plant photosynthesis and transpiration are substantially reduced. After several months of vigorous ET, surface water storage accessible to vegetation would be substantially depleted without sufficient recharge. As a result, an extended non-frozen period in autumn would have less impact on ET than a similar extension in the spring.

The results of the positive response of ET to an advancing and lengthening non-frozen season vary spatially and are even reversed in warmer, southern regions of the domain where water supply is a leading constraint for ET (Zhang et al. 2011). These results also are consistent with recent reports indicating increasing regional water stress (Angert et al. 2005; Schindler and Donahue 2006; Hogg et al. 2008) and associated vegetation browning and productivity reduction (Goetz et al. 2005; Zhang et al. 2009; Beck et al. 2011), higher tree mortality rates (Kurz et al. 2008; Mantgem et al. 2009), and increasing fire frequency (Westerling et al. 2006).

4.4 Regional/Basin ET and Impact on Water Balance

ET plays an important role in linking the water, energy, and carbon cycles and represents over 60% of precipitation over the global land area (Oki and Kanae 2006). A better understanding of recent changes in the Arctic regions requires linking ET with other hydrological components including precipitation and river discharge. Precipitation and river discharge measurements are currently available from Arctic observation networks (Yang et al. 2005; McClelland et al. 2004). Evapotranspiration (ET) is highly heterogeneous both spatially and temporally due to strong vegetation canopy control on transpiration. Relatively sparse measurements of these variables across the northern high latitudes make an accurate assessment of ET a challenge. Remotely sensed data, especially from polar-orbiting satellites, provide relatively frequent and spatially contiguous monitoring of surface biophysical variables affecting ET, including albedo, biome type, and vegetation density. Satellite-based ET products have been produced at regional and global scales with varying accuracy (Suzuki et al. 2018; Zhang et al. 2009; Cleugh et al. 2007; Mu et al. 2007; Fisher et al. 2008).

Zhang et al. (2009) developed an evapotranspiration (ET) algorithm driven by satellite remote sensing inputs, including AVHRR GIMMS NDVI, MODIS land cover, and NASA/GEWEX solar radiation and albedo, and regionally corrected NCEP/NCAR Reanalysis daily surface meteorology. The algorithm was used to assess spatial patterns and temporal trends in ET over the pan-Arctic basin and Alaska from 1983 to 2005 (Fig. 4.7). The annual ET patterns are spatially complex but show generally reduced ET with increasing latitude. The spatial pattern of ET also corresponds to distributions of the major biome types. Boreal forest regions have the highest annual ET ($277.35 \pm 68.89 \text{ mm yr}^{-1}$) among the three major biome types followed by grassland ($248.40 \pm 76.94 \text{ mm yr}^{-1}$) and Arctic tundra ($158.98 \pm 37.61 \text{ mm yr}^{-1}$). c water bodies resolved by the 1-km resolution global land cover classification cover only 6.4% of the pan-Arctic domain, they are distributed across a wide geographic range, with relatively large magnitude and spatial variation in annual ET ($410.35 \pm 142.58 \text{ mm yr}^{-1}$) relative to vegetated land areas. Meanwhile, both the GPCP and GPCC sources show similar multiyear mean annual precipitation patterns; though the GPCP precipitation rates are much larger, averaging 1.31 (± 0.38) times the GPCC precipitation rates. Both precipitation datasets

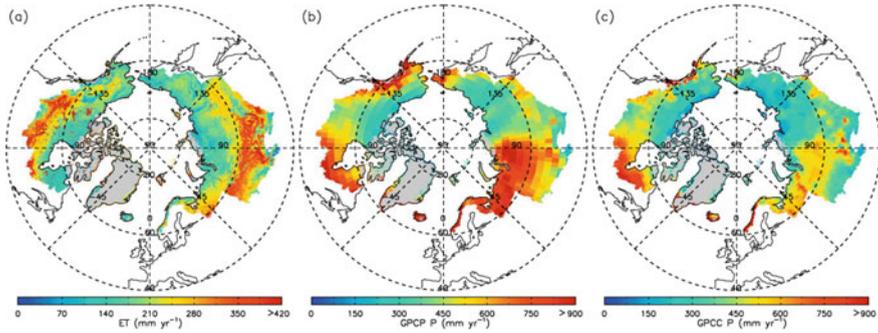


Fig. 4.7 Maps of multi-year (1983–2005) mean annual calculated ET **a** and precipitation derived from **b** GPCP and **c** GPCC sources (Zhang et al. 2009)

show portions of Southern Alaska, Northeastern Canada near Hudson Bay, and Western Eurasia having the largest precipitation, with relatively arid polar tundra areas showing the least precipitation.

The pan-Arctic region has a small positive trend in annual P for the 1983–2005 period (Fig. 4.8), indicated by both GPCP (7.4 mm decade⁻¹) and GPCC (4.3 mm decade⁻¹) sources, while ET shows a significant positive trend of 3.8 mm decade⁻¹ for this period coinciding with regional warming (Trenberth et al. 2007) and vegetation greening trends (e.g., Goetz et al. 2005; Zhang et al. 2008). On average, ET represents 44.4% (± 1.6) and 54.7% (± 1.8) of P for the pan-Arctic domain relative to the GPCP and GPCC sources, respectively. Both P and ET show positive trends so the net effect on P–ET is reduced and annual P–ET shows an insignificant wetting trend indicated by both GPCP and GPCC data. Both GPCP- and GPCC-derived results show P–ET interannual variability of approximately $\pm 6\%$ relative to the long-term mean. Time series plots of annual P–ET anomalies for the domain show that the years 1984–1985, 1995–1996, 1998–2001, and 2003 are dry relative to the long-term mean. Both regional average P and ET show seasonal changes during the 23-year period (Table 4.1). The GPCP and GPCC data show positive regional P trends in spring, summer, and autumn, with slightly decreasing P trends in winter. The ET results show significant ($P < 0.1$) positive trends for all four seasons, with the largest ET increases in spring and summer, similar to P. These results suggest that regional warming and associated lengthening of the seasonal non-frozen period are promoting increases in annual ET, while positive P trends during this period are counteracting these evaporative water losses and associated changes to the annual water balance. On a seasonal basis, larger positive trends in P relative to ET in spring, summer, and autumn imply that the pan-Arctic domain is becoming wetter during the growing season. In contrast, decreasing P and slightly positive ET trends indicate that the pan-Arctic domain is becoming drier during the winter period.

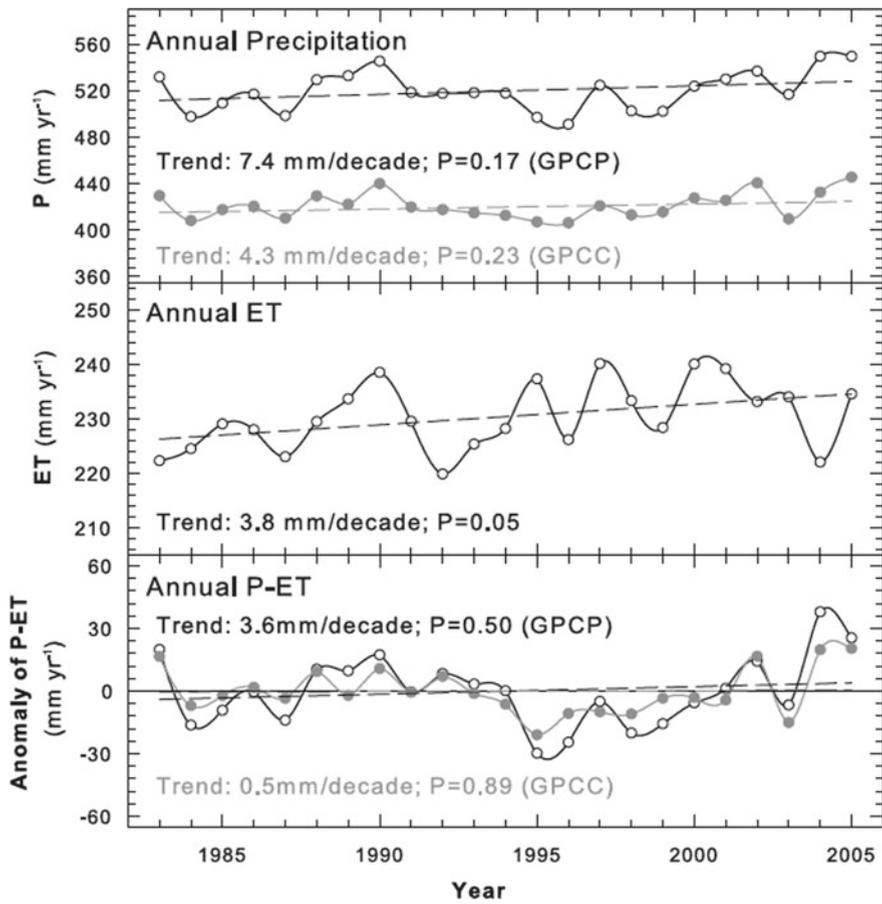


Fig. 4.8 Annual time series of precipitation (P) derived from GPCP and GPCCC data sources; ET derived from the NDVI-based ET algorithm, and corresponding annual $P-ET$ anomalies from the long-term (23-year) mean. Dashed lines show linear trends for the time series. The two P series from GPCP and GPCCC sources are highly correlated ($r = 0.91$; $p < 0.001$). The resulting $P-ET$ anomalies are also highly correlated ($r = 0.89$; $p < 0.001$). (Zhang et al. 2009)

Similar interannual changes were found by Suzuki et al. (2018) for the Mackenzie River, the Lena River, and the Yukon River basin. Comparisons were made among monthly precipitation (P) from GLDAS-2 and river runoff (R) for each river basin (Fig. 4.9). The maximum monthly P occurred in July for the Lena and Mackenzie River basins and in June or July for the Yukon River basin. The Lena River exhibits the largest amount of river runoff (R), followed by the Yukon River and the Mackenzie River. The peak value of R occurred in June for all three rivers. Using monthly climatology of river runoff data, the peak snowmelt occurs between April and June, accounting for approximately 40%, 36%, and 28% of the annual R

Table 4.1 Trends of annual precipitation, ET, and P–ET from 1983 to 2005 for the three primary regional biome types and North American and Eurasian portions of the pan-Arctic domain (Zhang et al. 2009)

Continent	Biome type	Area (10 ⁶ km ²)	P trend (mm de ⁻¹)		ET trend (mm de ⁻¹)	P-ET trend (mm de ⁻¹)	
			GPCP	GPCC		GPCP	GPCC
North American	Tundra	3.9	19.80**	8.77	2.81*	16.99*	5.97
	Forest	2.69	-7.97	-11.9*	-3.06	-4.90	-8.84
	Grassland	0.72	5.03	5.12	6.54*	-1.52	-1.42
Eurasia	Tundra	6.21	12.83**	8.70*	2.72	10.11*	5.98
	Forest	4.91	2.53	6.27	6.94**	-4.41*	-0.68
	Grassland	1.96	-9.99	-6.05	4.71*	-14.70	-10.77

* $p < 0.10$ ** $p < 0.05$

in the Lena River, Yukon River, and Mackenzie River basins, respectively. E increased steadily from 2002 to 2016 with slopes of 1.2–4.1 mm y⁻¹. The increase in E from GLDAS-2 dataset can explain most of the decrease in water storage. Thus, evapotranspiration driven by increasing summer temperatures may be the primary factor controlling water storage. This suggests that future warming might further decrease the water storage at high latitudes in the Arctic circumpolar region.

Evapotranspiration of the major Arctic river basins simulated by a land surface model, CHANGE (Park et al. 2011), shows larger interannual variability, with large regional differences (Fig. 4.10), representing differences in climate, soil, vegetation, and landscape. Interestingly, the interannual variability of ET is larger in relatively warm basins (i.e., Ob and Mackenzie) than cold (Yenisey and Lena). A considerable area of the latter is underlain by permafrost relative to the first. The recent warming temperature derives the warming of permafrost and thereby wetting soil that reduces moisture stress to ET. In other words, the wetted soil decreases the variability of ET under the warming temperature (Ohta et al. 2014). In reality, observations have identified the expansion of thermokarst lake resulted from permafrost degradation in the eastern Siberia (Ulrich et al. 2017), and consecutive positive anomalies of soil moisture were observed at the same region in the recent decade due to abnormally high winter precipitation (Iijima et al. 2010).

The four major basins show an increasing trend of ET over the period of 1979–2015 (Fig. 4.10). However, the trends evidently indicate large regional differences; the Siberian basins have significantly high increasing rates of ET (+0.72–+1.01 mm yr⁻¹) compared to Mackenzie. The significant increases in the Siberian basins are due to the recent warming temperature under the wetted soil moisture mentioned above because the temperature is a major deriving factor of ET (Park et al. 2008). The warming temperature also derives increases in vegetation biomass, so that the increased leaf area enhances transpiration and canopy interception to precipitation water. In contrast, the higher leaf area reduces solar radiation reaching

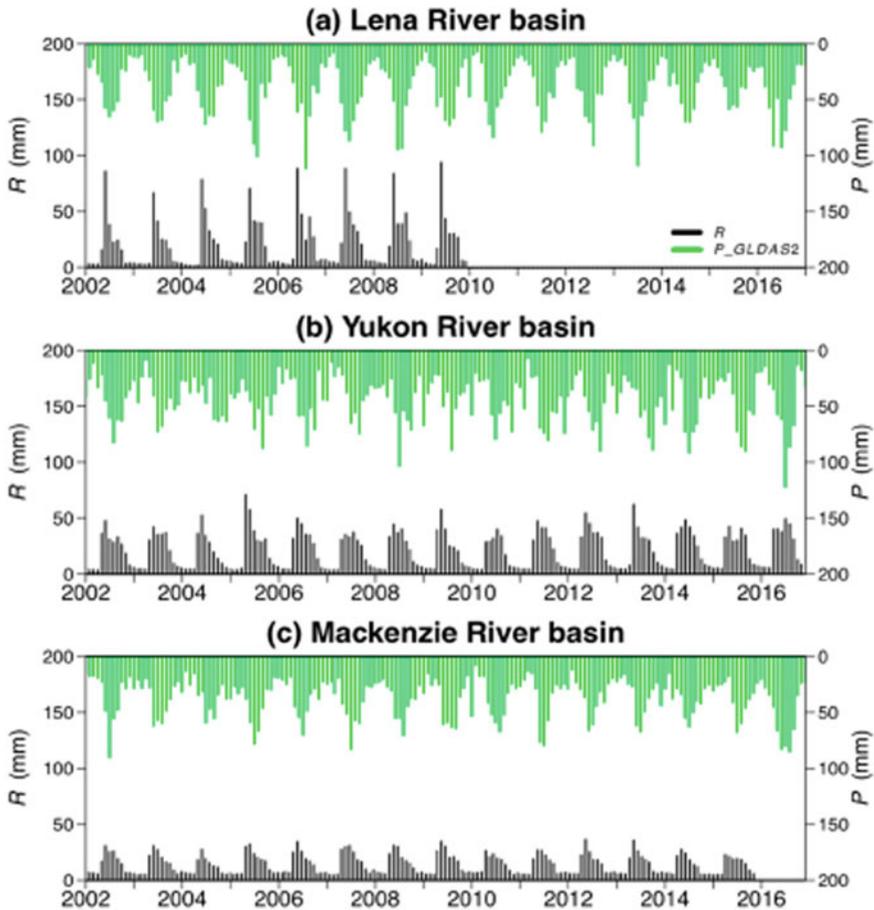
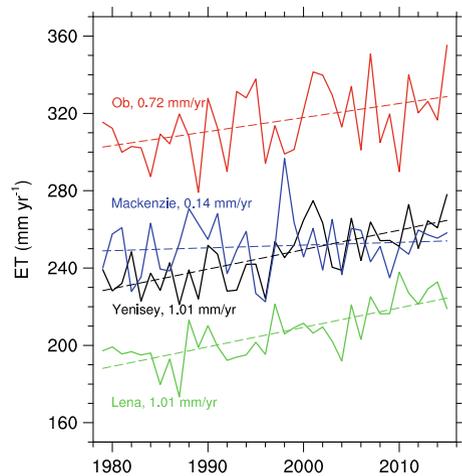


Fig. 4.9 Temporal variations in monthly water balance components from 2002 to 2016 in the **a** Lena River basin, **b** Yukon River basin, and **c** Mackenzie River basin (Suzuki et al. 2018, open access, no special permission is required to reuse all or part of the article published by MDPI, including figures and tables)

on soil surface and thereby less soil evaporation. However, most soil evaporation occurs in the spring season when solar radiation is the strongest, and leaf opening is not initiated (Park et al. 2008). In the Mackenzie River, warming was consistent with Siberia, while there was soil drought due to less precipitation during the recent period (Park et al. 2013). It suggests that the lower ET in the Mackenzie River for the recent decade is likely because of the influence of the drought. The comparison of ET between the river basins concludes that the differences in regional ETs mainly resulted from different climate and soil moisture conditions.

Fig. 4.10 Interannual variability of ET in the Arctic major river basins, simulated by a land surface model CHANGE using WATCH forcing dataset. The dot lines and numbers represent the ET trend in individual river basins



4.5 The Net Surface Moisture Flux (P-E)

Great hydrologic consequences are arising from the variations in the net surface moisture flux, which is the difference between precipitation (P) and evapotranspiration (E), over the large region and watersheds. For simplicity, here we use E to represent evapotranspiration over terrestrial surfaces and evaporation over ocean surfaces. In this respect, a key underlying question is: Will Arctic landscapes become wetter or drier as climate changes? Over the ocean, a corresponding question is: Will the ocean surface gain or lose freshwater to the atmosphere? While scaling issues complicate the answer to these questions, the fundamental uncertainty surrounds the surface moisture budget and the relationship P and E. If P (including both rain and snow) exceeds E over a period of time, the excess goes into runoff or storage. If E exceeds P, the surface moisture deficit leads to drying over land unless there is sufficient recharge from below. A drying surface leads to decreased water supplies, increased wildfire risk, and moisture stress on vegetation, all of which have consequences for terrestrial ecosystems and human activities. Over the ocean, the consequences of an imbalance of P and E will be a change of the near-surface stratification.

The trajectory of Arctic surface wetness is confounded by observations of decreasing soil moisture in the Arctic (Hinzman et al. 2013) and in subarctic Swedish basins where precipitation has been increasing (Destouni and Varrot 2014). Earth system model results also show a reduction in wetland extent in higher latitudes, largely associated with permafrost thaw (Avis et al. 2011). The most comprehensive assessment of trends of P-E based on historical data and model simulations appears to be that of Rawlins et al. (2010), who used a variety of precipitation datasets, atmospheric reanalyses, land surface model output, and global climate models. Rawlins et al. (2010) used various approaches, such as

average of nine global climate models (GCMs), average of five land surface models (LSMs), the surface energy balance and remote sensing (RS) method, the Variable Infiltration Capacity (VIC) model, and the ERA-Interim reanalysis, to synthesizing data, and found increases of E over Arctic terrestrial regions in recent decades. More generally, Rawlins et al. also found that trends of P, P-E, and river discharge were generally positive in the observational data, for which record lengths ranged from 20 to 50 years. However, trends of P-E, computed as differences between historical P datasets and satellite-derived (AVHRR GIMMS) E, showed no significant trend. The nine global climate models examined by Rawlins et al. (2010) showed statistically significant trends of terrestrial pan-Arctic P-E over the period 1950–1999 in eight of the nine cases, and in all nine cases for the period 1950–2049. Trends for the historical period were smaller than for the future period in the climate model output. All results were for annual means. The positive trends in annual mean P-E contrast with the expectation that longer and warmer summers will increase E sufficiently to favor summer drying. Anticipated increases of high-latitude wildfire activity (Partain et al. 2016; Flannigan et al. 2015) are consistent with this expectation, highlighting the mixed picture of future surface wetness trends in the Arctic. For the Arctic marine areas, Bintanja and Selton's (2014) analysis of model output found that increased open water drove an increase of E that, in turn, was a major contributor to future increases of P over the Arctic Ocean in climate model simulations. However, compelling demonstrations of increases of E over the Arctic Ocean based on historical data are generally lacking.

For the future, one of the earlier examinations of pan-Arctic hydrology changes projected by climate models found that Arctic P-E increased in model simulations of the twenty-first century (Kattsov et al. 2007), implying a wetter Arctic surface in the future. However, that study presented changes in only the annual pan-Arctic P-E and did not consider differences between ocean and land areas nor between different terrestrial subregions. Model-based studies summarized below are unanimous in projecting future increases of river discharge in the Arctic. The coarse resolution of these models and their rudimentary treatment of permafrost and vegetative processes make it questionable to base conclusions about Arctic surface wetness trends on results from these simulations. A more recent evaluation of 25 CMIP5 global climate model projections, although global rather than the Arctic in scope, distinguished summer and winter changes of E (Laine et al. 2014). The results presented by Laine et al. (2014) highlight the challenge of assessing future changes in high-latitude surface wetness. As shown in Fig. 4.11 (Laine et al. 2014), The projected changes of P and E show the expected pattern, with increases of E over Arctic land areas, primarily during summer, and strong increases over the subarctic seas during winter in areas of sea ice loss (consistent with Bintanja and Selten 2014). However, the middle panels of Fig. 4.11 show that changes of P-E over the Arctic are much less spatially coherent and less robust than changes of P and E separately. In particular, summer P-E is projected to decrease much of Canada, increase over Alaska, decrease over the western and northern Eurasian subarctic, and increase over parts of northeastern Russia, including Chukotka. Over most of these areas, the sign of the projected change is not robust across the models at the

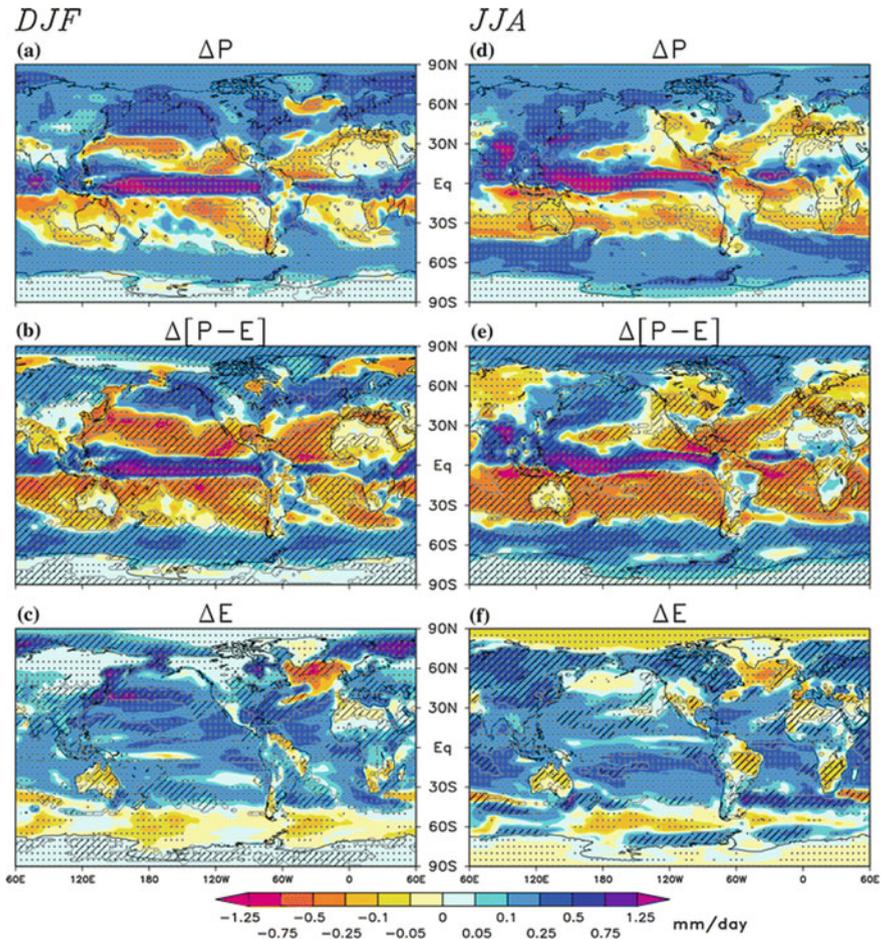


Fig. 4.11 25-model mean changes in precipitation P (upper panels), $P-E$ (middle panels), and evapotranspiration E (lower panels) for Dec–Feb (left panels) and Jun–Aug (right panels). In dotted regions, the sign of change is robust among the different models at the 95% confidence level. Hatching in middle and lower panels indicates where a change in $P-E$ is dominated by P and E , respectively (Laine et al. 2014, with open access)

95% confidence level. Figure 4.11e’s spatial pattern over high-latitude land areas is very consistent with the projected changes in soil moisture obtained by Dirmeyer et al. (2013) using 15 of the same models. Dirmeyer et al. (2013) showed reductions of summer soil moisture over northern Canada and north-central Russia but not over Alaska and eastern Siberia, from the preindustrial to the twentieth century. However, decreases of summer soil moisture do spread to southern and central

Alaska in the twenty-first century (Dirmeyer et al. 2013). The results of these two studies highlight the uncertainty in the trajectory of surface wetness in Arctic land areas in the present generation of global climate models.

River discharge data, which can be viewed as proxies for P-E integrated over river basins, generally show increases from the mid-twentieth century to the early 2000s, with varying degrees of statistical significance (Yang et al. 2004a, b; Shiklomanov and Lammers 2009; Overeem and Syvitski 2010; Holmes et al. 2013; Bring and Destouni 2014). Model-based studies consistently show that projected changes in atmospheric forcing will drive twenty-first-century increases in high-latitude river discharge (e.g., Vliet et al. 2013; Koirala et al. 2014; Bring et al. 2015).

Various factors contribute to the uncertainty in the projected changes in Arctic surface wetness. Internal variability, which can affect trends over decadal and multidecadal timescales, is clearly one consideration. However, studies such as those of Rawlins et al. (2010), Laine et al. (2014) and Dirmeyer et al. (2013) have utilized multimodel ensembles that tend to average out internal variations, and these studies indeed contained robust signals in the future changes of P and E, although not the hydrologically critical difference, P-E. Across-model differences in process formulations, especially the formulations in the terrestrial modules, almost certainly are key contributors to the uncertainty. Because of their coarse ($\sim 100\text{--}200$ km) resolution, global climate models are often unable to resolve terrain and vegetation variations that characterize the Arctic landscape. The terrestrial modules of many global models include crude treatments of permafrost, with an inadequate representation of soil horizons and resolution of the changing active layer. Important differences in Arctic vegetation such as tundra types (e.g., heath vs. tussock), forest composition, and effects of small lakes and ponds are generally not well considered. Also not included in most global models are local topography, subgrid-scale vegetation distributions, and permafrost hydrology, which are all important determinants of soil moisture, drainage, and, to at least some extent, land surface evaporation.

4.6 Conclusion and Discussion

Evapotranspiration (ET) is a key component in global water and energy cycles. In the northern regions where water is either permanently or seasonally frozen, ET is influenced by both hydrological and thermal conditions of the land surface through complex physical processes and feedbacks to the surrounded environment. This chapter presents and discusses recent research advances about the ET in northern regions. ET in northern regions tends to increase with the decrease of latitude. The largest ET typically appears in the forest ecosystem, while the grasslands and shrublands have the small ET. While the seasonal variations in ET are usually high,

the interannual variability in annual ET is usually low over the Arctic regions, reflecting the dominance of cold temperature constraints on northern ecosystem processes. It should be noted that the sublimation from snow cover accounts for about 15% (in Siberia) to 25% (in western Canada) of winter precipitation. This is certainly a non-negligible hydrological component that affects the end of winter snow mass, river discharge, and regional water resources across the broad northern regions. Many factors, such as soil moisture, vegetation type, and productivity, and ecosystem features affect ET over the northern regions. In addition, precipitation plays a key role in impacting ET. ET is more sensitive to precipitation in the early growing season than in the late growing season. Furthermore, changes in freeze–thaw processes due to warming also affect land surface conditions and the ET processes. Such an effect is different between the southern and northern parts of the Arctic because of the water supply in the soil. Therefore, annual ET shows positive correlations with the timing of primary seasonal thaw in the northern Arctic region, in which soil water is relatively large. This is because the earlier onset of springtime thaw facilitates earlier photosynthesis and transpiration and generally enhances plant growth and transpiration when plant-available moisture is high. However, the reverse relationship is witnessed in the southern Arctic regions where the water supply is usually limited. From 1983 to 2005, ET increased significantly in the Arctic region with a rate of $3.8 \text{ mm decade}^{-1}$ because of regional warming and vegetation greening. Such an increase in ET may exert significant impacts on the regional hydrology and water resources. Advanced models can simulate past ET change over the large northern watersheds. Remote sensing has proved new ET data and information that support climate and hydrology research and applications. It should be noted that most available ET studies in the Arctic region rely heavily on the models and/or satellite observations, while the ground measurements of ET in such a harsh but hydroclimatically important regions are yet extremely sparse. It is therefore highly recommended that the community should focus more on improving our understanding of ET processes and its feedbacks to climate by integrating a wide range of in situ observations and physical-based modeling approach. Finally, it is a key underlying question: Will Arctic landscapes become wetter or drier as climate changes? Global models suggest increases of E over Arctic terrestrial regions in recent decades. Rawlins et al. also found that trends of P, P-E, and river discharge were generally positive in the observational data. For the future, summer P-E is projected to decrease much of Canada, increase over Alaska, decrease over the western and northern Eurasian subarctic, and increase over parts of northeastern Russia. Over most of these areas, the sign of the projected change is not robust across the models at the 95% confidence level. Many factors contribute to the uncertainty in the projected changes in Arctic surface wetness. There is certainly a need to quantify and narrow the uncertainties in global models over the large northern regions.

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Yinsheng Zhang is a Professor of Hydrology and Physical Geography in the Institute of Tibetan Plateau Research, Chinese Academy of Sciences (ITPCAS). He received his Ph.D. with a major in physical geography from Lanzhou Institute of Glaciology and Geocryology, Chinese Academy of Sciences in 1994. Prior to joining ITPCAS in 2009, He was a Senior Research Scientist at the Institute of Observational Research for Global Change, Japan Agency for Marine-Earth Science and Technology for more than 10 years. He is particularly interested in land surface hydrological processes in the cold regions and has carried out extensive fieldwork to monitor the hydrological processes in the Tibetan Plateau, Eastern Eurasia, Mongolia, etc. He has been the PI for six research projects from the National Natural Sciences Foundation of China (NSFC), Ministry of Science and Technology (MOST) of China, and Chinese Academy of Sciences. He has authored and co-authored more than 90 papers in peer-reviewed journals and book chapters. He was awarded as “Excellent Graduate Advisor of Chinese Academy of Sciences” in 2018.



Ning Ma is a Postdoc Researcher at the Institute of Tibetan Plateau Research, Chinese Academy of Sciences. He received his B.S. with a major of Geographical Sciences and M.S. with a major of Earth System Sciences from the Department of Earth and Environmental Sciences, Lanzhou University in June 2010 and December 2012, respectively. He received his Ph.D. from the Institute of Tibetan Plateau Research, Chinese Academy of Sciences in June 2017. His research interests include land surface processes, hydrological processes, and climate change in cold and arid regions. He has authored and co-authored more than 30 publications in peer-reviewed journals and he was also selected as an award winner of “Hundred Excellent PhD theses of Chinese Academy of Sciences” in 2018. The overall goal of his research is to improve our systematic understanding of the evapotranspiration processes and its response to climate and land cover changes through applications of an integrated research approach consisting of field data analysis and process-based modeling.



Dr. Hotaek Park is a Senior Scientist at JAMSTEC. He earned his Ph.D. at Graduate School of Bioagricultural Sciences, Nagoya University in 2000. Then, he worked at two Japanese Institutes as postdoctoral fellowship and joined the position of JAMSTEC research scientist in 2007. He is an expert on hydrology, biogeochemistry, and climate researches in cold regions, with interests in evaluating changes in land surface processes under climate changes and predicting future changes using land surface model, remote sensing data, reanalysis products, in site observations, and model outputs. He has the author/co-author of numerous scientific articles that have published on a variety of peer-reviewed journals, conference proceedings, and books. His recent research is focused on assessing impacts of changing snow and permafrost on hydrological processes in the context of climate variability and interactions between declining Arctic sea ice and terrestrial ecohydrologic processes coupling models of land-atmosphere-ocean processes.



Dr. John E. Walsh is the Chief Scientist of the International Research Center and President's Professor of Global Change at the University of Alaska, Fairbanks, USA. He is also an emeritus faculty member of the University of Illinois. He has been involved in Arctic research and education for more than 40 years. His more recent research has addressed Arctic climate and weather variability, with an emphasis on the drivers of cryospheric variations over seasonal to multidecadal timescales. He has served as lead author for assessment reports of the Intergovernmental Panel on Climate Change, the Arctic Monitoring and Assessment Programme, and the U.S. National Climate Assessment. He has also co-authored a textbook, *Severe and Hazardous Weather*. He received a B.S. in Mathematics from Dartmouth College and a Ph.D. in Meteorology from the Massachusetts Institute of Technology.



Ke Zhang is a Professor of Hydrology and Water Resources in College of Hydrology and Water Resources at Hohai University, China. He received his Ph.D. in Ecohydrology in 2009 from University of Montana, USA and his Master's in Hydrology and Water Resources from Hohai University, China. Prior to joining Hohai University, he worked as a research scientist and research associate at University of Oklahoma and Harvard University for five years. His research interest includes hydrological modeling, hydrometeorology, ecohydrology, land surface remote sensing and modeling, and climate change. He has published more than 90 articles in peer-reviewed journals, authored 6 book chapters, and presented more than 60 research papers in national and international conferences. He has frequently served as reviewers for more than more than 30 international journals. He also serves as associate editor for Journal of Hydrology, Frontiers in Earth Science, and Water Science and Engineering, and editorial board member for Environmental Modelling & Software, Scientific Reports, and Hydrology Research.