

Chapter 13

Hydrological Aspects in a Siberian Larch Forest

T. Ohta

13.1 Introduction

Understanding the water cycle is crucial for ecological studies, because all life both depends on water and affects the water cycle. Energy exchanges between biomes and the atmosphere are also strongly tied to water cycles through evapotranspiration. Therefore, clarifying hydrological processes and features can greatly contribute to a deep understanding of ecological systems.

The high latitudes are characterized by low precipitation and short growing seasons due to the cold winter climate. Several studies have used general circulation models (GCMs) to examine the effects of high-latitude forests on water, energy, and carbon (WEC) cycles (e.g., Bonan et al. 1992, 1995) and have been followed by in situ observational studies. Some projects of WEC cycles have been carried out in high-latitude regions, including the northern hemisphere climate processes land surface experiment (NOPEX) on the Scandinavian Peninsula (Halldin et al. 1998) and the Boreal ecosystem-atmosphere study (BOREAS) in North America (Sellers et al. 1995). In Siberia, WEC cycles observations began in the late 1990s. The terrestrial carbon observation system-Siberia (TCOS-Siberia) study examined WEC exchanges in European Russia and Western Siberia (Schulze et al. 2002). Through intensive observations and analyses, the study revealed numerous characteristics not only of grasslands and bogs but also of forests, as well as distinctive characteristics of high-latitude WEC cycles. Examples of high-latitude WEC characteristics include relatively low evapotranspiration, the importance of tree phenology, and large contributions of understory vegetation.

Despite its having a more severe climate than the regions mentioned above, much of Northeastern Siberia is covered by taiga forests dominated by *Larix cajanderi* (Chap. 3, this Vol.). Yakutsk, located along the middle reaches of the Lena River, has an annual average temperature of only -10.2°C and annual precipitation of 235 mm. Moreover, extensive continuous permafrost exists in Northeastern Siberia, while relatively little permafrost exists in North America, Scandinavia, and Western Siberia (see Fig. 1.1. this Vol.). Therefore, the interactions between forests and permafrost in Northeastern Siberia likely result in unique hydrological features

in this region compared with other climate zones. As a first attempt to understand WEC cycles in Northeastern Siberia, Kelliher et al. (1997) conducted field studies for a short period during summer 1993 using an eddy correlation method. Long-term studies near Yakutsk have subsequently been carried out, including collaborative research between Russia and Japan as part of the global water and energy cycle experiment (GEWEX), Asian Monsoon Experiment-Siberia (GAME-Siberia; Lawford et al. 2004; Yasunari 2007), the Core Research for Evolutional Science and Technology (CREST)/Water and Energy Cycles in Northern Forests (WECNoF) project (<http://www.agr.nagoya-u.ac.jp/~wecnof/>), and the Russian–Dutch project PIN-Matra (Moors et al. 2006).

This chapter first discusses the hydrological characteristics of a Northeastern Siberian larch forest and then examines hydrological differences between permafrost and non-permafrost regions. Finally, the water cycle is estimated at a continental watershed scale using a land surface model (LSM) that incorporates in situ observations.

The water and energy balances are strongly related to the carbon dioxide exchange. Net ecosystem exchange of carbon dioxide in a permafrost larch stand is discussed by Y. Nakai (see Chap. 10, this Vol.).

13.2 Approaches to Study Stand-scale Hydrological Characteristics in a Larch Forest of Northeastern Siberia

13.2.1 Study Site for the Stand-scale Investigation

The observation site is located at the middle reaches of the Lena River, approximately 20 km north of Yakutsk city (62°N, 129°E, 220 m a.s.l.), and on the left bank of the river. This site, called Spasskaya Pad, is part of an experimental forest of the Institute of Biological Problems in the Cryolithozone, Russian Academy of Sciences. A 32-m-high meteorological tower was installed at the site in 1996, and full observations have been carried out since 1998. The topography is flat but inclining slightly northward.

The dominant overstory species is *Larix cajanderi*; the larch stand has a density of 840 trees ha⁻¹, a mean tree height of 18 m, and a leaf area index (LAI) of 1.56 m²m⁻². The forest floor is fully covered with *Vaccinium vitis-idaea*, with an LAI of approximately 2.0 m²m⁻². Roots are concentrated in a surface soil layer 0–20 cm deep; only 1.5% of the surface area of roots is distributed below 50 cm deep. *Larix cajanderi* is a deciduous species having clear phenological changes throughout a year. From 1998 to 2006, larch leaves started to flush from late April to late May (weeks of the year 16–21), and the canopy reached a mature condition during the following 2–4 weeks. The start of leaf fall varied yearly but generally occurred from late August to late September (weeks of the year 34–37), and almost all leaves had fallen by 10–20 days after the start of leaf fall.

Further details of the site have been provided by Ohta et al. (2001), Dolman et al. (2004), Hamada et al. (2004), and Ohta et al. (2008).

13.2.2 *Measurement of Meteorological and Environmental Variables*

Meteorological elements were measured at several heights on the 32-m-high tower, at six levels from 1998–2000 and at four levels since 2001. Downward and upward shortwave radiation (S_d , S_u) and downward and upward longwave radiation (L_d , L_u) were observed at the top of the tower and above the forest floor. In addition, photosynthetically active radiation (PAR), both downward and upward (P_d , P_u), was measured at two levels. Wind direction was observed at the top of the tower only, while wind speed was observed at six or four levels using three-cup anemometers. Air temperature and relative humidity were measured at all levels, with ventilation.

Soil temperature (T_g) and soil water content (SWC) were also measured at several depths. Platinum resistance thermo-sensors were used to measure soil temperature, and time domain reflectometry (TDR) soil-moisture sensors were used to obtain SWC. Ohta et al. (2001) and Ohta et al. (2008) have reported details of these measurements.

13.2.3 *Measurement of Water Vapor and Energy Fluxes*

Water vapor and energy fluxes above the canopy were obtained by an eddy correlation method using a three-dimensional ultrasonic anemometer (3D-SAT) and an infrared gas analyzer (IRGA). Energy balance above a canopy was given as follows:

$$R_n - G = H + LE + J_a + J_s + J_w, \quad (13.1)$$

where R_n is net all-wave radiation (W m^{-2}), G is ground heat flow (W m^{-2}), H is sensible heat flux (W m^{-2}), LE is latent heat flux (W m^{-2}), and J_a , J_s , J_w are heat storage in the air mass below the 3D-SAT and IRGA, in the soil layer between the soil surface and ground heat flow plate, and in the biomass. The J_a consists of sensible heat storage (J_h) and latent heat storage (J_l).

The relationship between available energy and the sum of turbulent fluxes and heat storage is given as

$$H + LE + J_a + J_s + J_w = \alpha(R_n - G) + \beta, \quad (13.2)$$

where α , β are coefficients. Energy balance is completely closed when α is 1.0 and β is 0.0. At the study site, α and β have values of 0.90 and -4.0 , respectively. The value of α dropped to 0.75 when heat storages J_a , J_s , and J_w were ignored at the site (Ohta et al. 2001; Matsumoto et al. 2008a). The magnitudes of J_h and J_l were not small but led to small J_a , because these storages often had reverse directions and offset each other. This result implies that heat storage in the surface soil and biomass

is very important for energy balance in a Northeastern Siberian larch forest, because a sparse larch forest allows more radiation to penetrate under the canopy.

13.2.4 Evaluation of Hydrological Cycles in the Lena River Basin

Hydrological aspects at a continental watershed scale were investigated using an LSM and a distributed runoff model (DRM) coupled with the LSM. These models were applied over Central and Northeastern Siberia, including the Lena River and Kolyma River basins.

The LSM used was originally proposed by Yamazaki (2001) and developed to a watershed scale by Park et al. (2008). The model included three submodels for vegetation, snow, and soil processes. In the vegetation submodel, the canopy was divided into crown space and trunk space. The physiological responses to environmental variables were described by a Jarvis-type model. The parameters in the conductance responses are universal for all vegetation types in Far East Asia, from boreal to warm-temperate forests, based on the results obtained by Matsumoto et al. (2008b). Permafrost processes (i.e., melting and refreezing of the soil layer) were performed in the soil submodel. The watersheds analyzed were divided into $0.5 \times 0.5^\circ$ grids, and outputs from the LSM were obtained in each grid.

The DRM was originally described by Lu et al. (1989) and developed for Siberian rivers by Hatta et al. (2009). The runoff processes on slope and river channels were calculated in this model using a kinematic wave method. Hydraulic characteristics of channels were obtained from satellite data. Input for this DRM was outflow from a soil column simulated by the LSM described above. The DRM described not only the runoff processes but also breaking-up and freezing processes of river ice.

13.3 Seasonal and Interannual Variation of Energy Partitioning above the Siberian Larch Forest

Here, we discuss the partitioning characteristics of available radiation energy, $R_n - G$, into latent and sensible heat fluxes on seasonal and interannual temporal scales. Figure 13.1 shows the seasonal variation of net all-wave radiation; ground heat flow, including the energy for snowmelt; latent and sensible heat fluxes; and the Bowen ratio (the ratio of latent heat flux to sensible heat flux) above the canopy in 1998 based on daily values (Ohta et al. 2001). It shows only time series obtained over a dry canopy condition. The bar at the bottom of Fig. 13.1 shows the seasonal change of land surface conditions such as snow coverage and leaf phenology, which was determined by in situ observation.

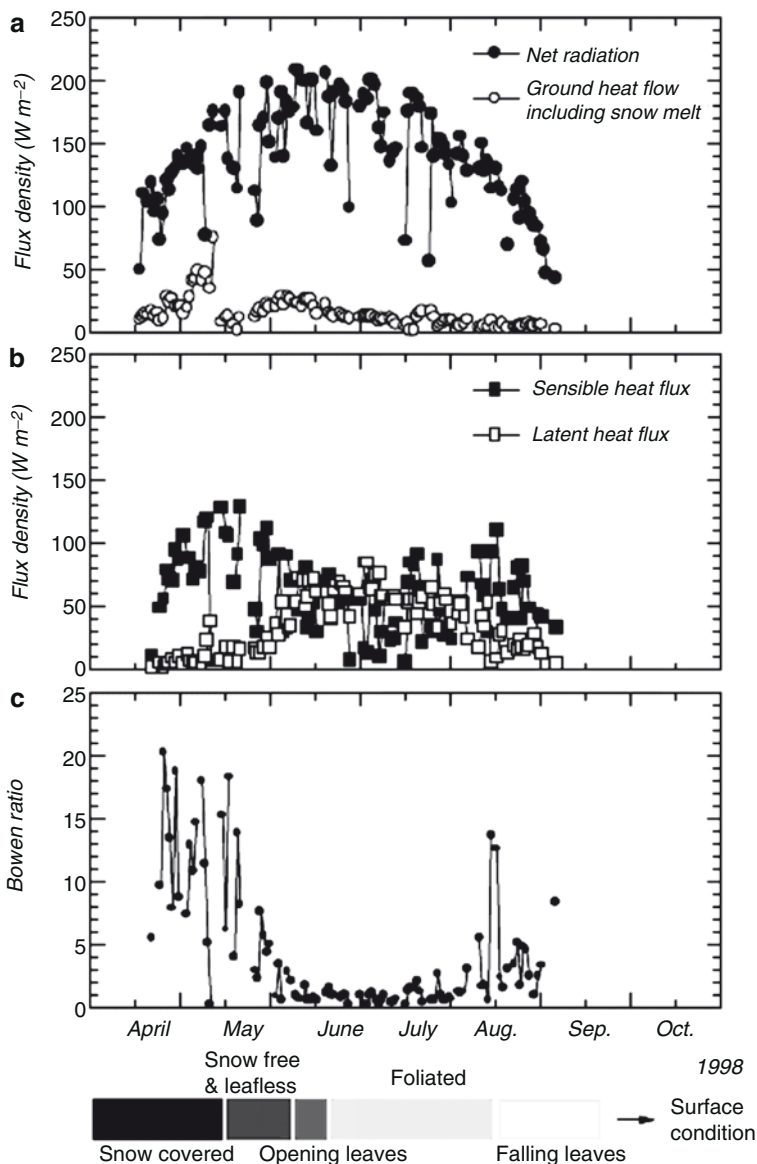


Fig. 13.1 Seasonal variation of energy budget components, Bowen ratios and land surface conditions obtained in the Siberian larch forest in 1998. (a) Net all-wave radiation (*filled circle*) and ground heat flow (*open circle*); (b) sensible heat flux (*filled square*) and latent heat flux (*open square*); and (c) Bowen ratios. (Ohta et al. 2001)

The latent heat flux remained low until the beginning of leaf emergence (i.e., bud break) at the end of May and then increased quickly toward the middle of June as the leaves opened and developed fully. The timing of leaf emergence of the larch trees almost agreed with the timing of permafrost thawing to depths of 10–20 cm. On the other hand, the sensible heat flux peaked in the middle of May and then dropped suddenly, even though net all-wave radiation was still increasing. Consequently, the time series of Bowen ratios (Fig. 13.1c) had a clear U-shape, with a minimum value of 1.0 during the growing season (June and July). This result indicates that the phenological effect on water and energy exchanges above the canopy is more significant than that of snow disappearance, although the soil surface condition became quite wet just after snow melt. Snow melt is a significant hydrological event from the viewpoint of runoff processes, but it does not have remarkable effects on the water and energy exchanges between forests and the atmosphere. Further, the minimum Bowen ratio (~1.0) was higher than that obtained at tropical and temperate forests, but typical for or somewhat lower than values at other boreal forests (e.g., Sellers et al. 1995; Kelliher et al. 1997, 1998; Tchebakova et al. 2002).

An evaporative fraction, presented by Barr et al. (2006), was used to examine the interannual energy balance features:

$$\Lambda = \frac{LE}{LE + H}, \quad (13.3)$$

where Λ is the evaporation fraction. Figure 13.2 shows the interannual time series of the latent and sensible heat fluxes above the canopy (upper panel) and the evaporative fraction (lower panel) (Ohta et al. 2008). Interannual fluctuation of latent heat flux was smaller than that of sensible heat flux. The evaporative fraction ranged from 0.379 to 0.678, with an average value of 0.477 and standard deviation (SD) of 0.182 during the growing season (June, July, and August). The corresponding Bowen ratios for these months were 1.10, 1.64, and 0.475, respectively.

As mentioned above, the average evaporative fraction was smaller in the Siberian larch forest than in temperate forests. Hamada et al. (2004) obtained similar Bowen ratios (1.0–2.0) in a Scots pine (*Pinus sylvestris*) forest located about 2 km south-east of the present study site of *L. cajanderi* forest during mid-summer of 2000. A small portion of available energy was partitioned to latent heat flux in the Siberian forest. However, the highest evaporative fraction was 0.678, which is equal to or somewhat larger than the values reported for several temperate forests (Kosugi et al. 2006, 2007). High evaporative fractions were found in wet years. These results imply that the low partitioning rate into latent heat flux is not an inherent characteristic of Siberian larch forest and that the portion of latent heat flux will increase under the more suitable growing conditions (e.g., higher soil water condition).

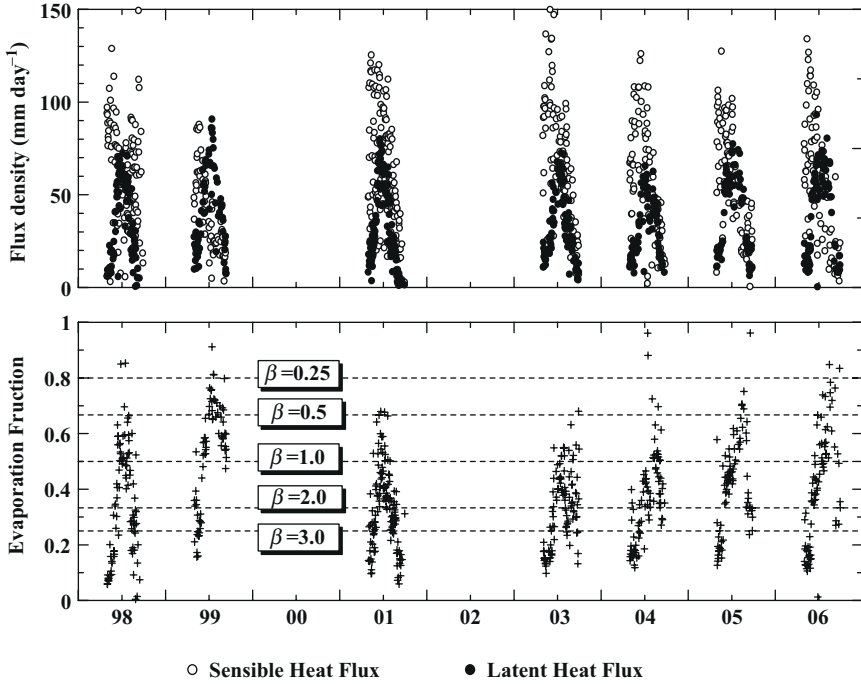


Fig. 13.2 Interannual variation of sensible heat flux (*open circle*) and latent heat flux (*filled circle*) (*upper panel*), and evaporative fraction (*lower panel*) over the Siberian larch forest during a period 1998–2006 (Ohta et al. 2008)

13.4 Water Balance of One-dimensional Scale in the Siberian Larch Forest

13.4.1 Interannual Variation

The water balance is described as follows:

$$P = E + \Delta S + R, \quad (13.4)$$

where P is the amount of precipitation (mm), E is the amount of evapotranspiration (mm), ΔS is the change in soil water storage between the first and the last day in a water budget period (mm), and R is the residual (mm). Positive R indicates runoff from a water balance system, while negative R indicates inflow into the system. In this section, because we focus on the temporal variation of the water balance in the Northeastern Siberian larch forest, the unit of each term in Eq.(13.4) is mm y^{-1} . Here, we define a water year as the 365 days from October 1 of the previous calendar

year to September 30 of the current year. That is, year “N” means the 365 days from October 1 of year “N-1” to September 30 of year “N.”

Figure 13.3 shows the interannual variation of each water balance component presented in Eq.(13.4) (Ohta et al. 2008), and Table 13.1 presents the means and SDs of the components (Ohta et al. 2008). Evapotranspiration in the cold period from October to April is not included in the figure and the table because of the negative net all-wave radiation and very low temperatures. The annual precipitation amount ranged widely from 111 to 347 mm y^{-1} (SD=82.3 mm). In comparison, the annual amount of evapotranspiration was relatively steady (SD=19.7 mm), varying from 169 to 220 mm y^{-1} .

In non-permafrost regions with low annual precipitation (particularly areas with less than 500 mm y^{-1} of precipitation), the annual amount of evapotranspiration has been found to vary linearly with the annual precipitation amount, with a very slight deviation (Zhang et al. 2001). Figure 13.4 shows the empirical relationships between annual precipitation and annual evapotranspiration in watersheds covered

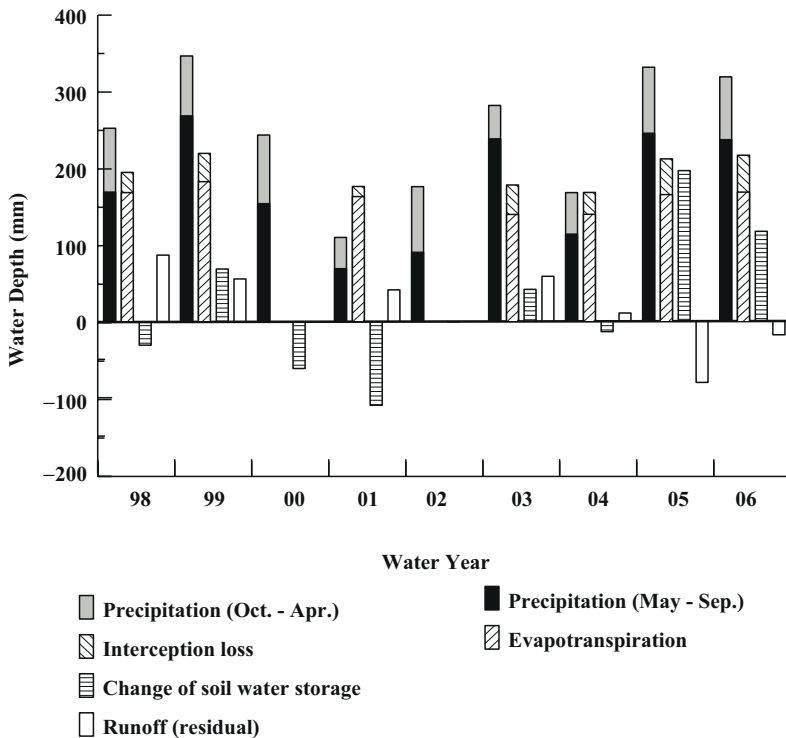


Fig. 13.3 Interannual variation of water balance components in the Siberian larch forest during a period 1998–2006. One water year is defined as 365 days from October 1 in the previous year to September 30 in the following year. (Ohta et al. 2008)

Table 13.1 Summary of the water balance in the Siberian larch forest during the period 1998–2006 (*left column*), 1998–2003 (*center column*), and 2004–2006 (*right column*) (Ohta et al. 2008)

	1998–2006		1998–2003		2004–2006	
	Mean	Standard deviation (SD)	Mean	SD	Mean	SD
Precipitation (mm)	259.2	82.3	248.4	86.3	273.6	78.4
October–April	69.0	18.3	63.7	19.3	76	14.2
May–September	190.3	70.1	184.7	76.8	197.6	59.9
Evapotranspiration (mm)	196.1	19.7	193.1	17.3	200.2	21.9
Change of soil water storage (mm)	39.6	94.1	–6.6	69.6	101.3	87.0
Residual (mm)	23.8	52.8	61.9	16.4	–27.0	38.4
Maximum thawing depth of permafrost (cm)	147.0	27.6	127.0	7.7	182.7	16.3
Date of maximum thawing depth	252.6	11.4	245.5	9.8	262.0	4.6

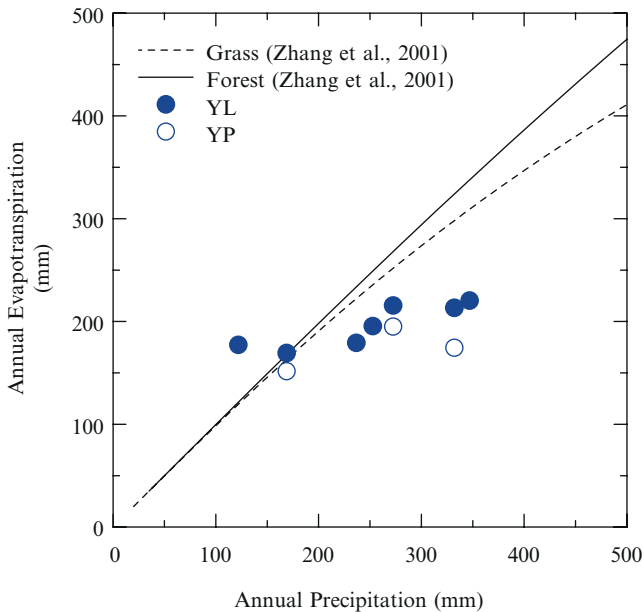


Fig. 13.4 Relationships between annual precipitation and evapotranspiration in the Siberian larch forest (*filled circle*) and pine forest (*open circle*). The solid and dashed lines indicate estimations by Zhang et al. (2001) for forested watersheds and grassland watersheds, respectively, located in a non-permafrost region. (Original data of larch and pine forests are from Ohta et al. 2008 and Hamada et al. 2004, respectively.).

with forests (solid line) and grass (dotted line) with an annual precipitation range of 0–500 mm y^{-1} , as obtained by Zhang et al. (2001). This figure also presents results for the larch forest (closed circles) and the nearby Scots pine forest (open circles) located approximately 2 km southeast of the larch forest (Hamada et al. 2004). Variation of evapotranspiration was much smaller in the Siberian forest ecosystem than in non-permafrost forests. This contrast is very important in understanding the effects of permafrost on hydrological aspects in this region.

Sugimoto et al. (2002) examined the water usage of larch stands near Yaktsuk using a stable isotope of oxygen, δO^{18} . They took water samples from precipitation, soil layers at several depths, and trees. The results showed that trees uptook precipitation water in wet years, but transpired permafrost meltwater in drought years. This result indicates that permafrost serves as a buffer for interannual fluctuation of precipitation.

As shown in Table 13.1, there was a significant difference in changes in soil water storage ΔS and residual R and in the maximum thawing depth of permafrost between the periods 1998–2003 and 2004–2006. The change of SWC was very small, -6.6 mm, with a wide SD of 69.6 mm during 1998–2003. The residual indicated runoff of 61.9 mm during this period. By contrast, in 2004–2006 the changes of SWC significantly increased to 101.3 mm and the residual became a negative value of -27.0 mm. In addition, the thawing depth of permafrost was much deeper

in 2004–2006 than in 1998–2003. The negative residual indicates inflow into the water balance system and implies that the increase in SWC cannot be explained only by the approximately 25 mm increase in precipitation. Ohta et al. (2008) discussed several possible reasons for the negative residual, but the exact causes have not yet been determined. Regardless of the reasons, the soil water regime is a key factor in understanding hydrological features in the permafrost regions.

13.4.2 Annual Evapotranspiration and Environmental Variables

Here, we discuss environmental parameters affecting the interannual variation of evapotranspiration at the Siberian larch forest. Several previous reports have examined environmental variables affecting evapotranspiration on diurnal and seasonal scales in Siberian forests (Ohta et al. 2001; Hamada et al. 2004; Dolman et al. 2004). These studies have noted the importance of atmospheric water deficit in Northeastern Siberian larch and pine forests at these temporal scales. The variables affecting WEC cycles likely vary according to the spatiotemporal scale of focus (Baldocchi et al. 2001; Katul et al. 2001; Stoy et al. 2005). However, little is known about the environmental parameters affecting evapotranspiration at an interannual scale because of the lack of long-term data for permafrost regions.

Figure 13.5 shows the interannual variations of four environmental variables (short-wave radiation, air temperature, atmospheric water vapor deficit, and SWC) and evapotranspiration normalized by potential evaporation E/E_p , and their relationships. These values were averaged for the growing season (June, July, and August). No significant relationships were observed except for evapotranspiration and surface soil moisture content. It might seem surprising that we did not find a clear relationship between mean temperatures and E/E_p , because air temperature is considered to be one of the most important variables limiting evaporation in cold regions. Dolman et al. (2004) emphasized the importance of surface conductance for daily or seasonal variation of evapotranspiration; they reported that evapotranspiration was low when trees were exposed to very dry air, even if soil moisture did not limit stomata closure. However, we found no clear relationship between atmospheric water vapor pressure and E/E_p , as shown in Fig. 13.5c. These results basically agree with those obtained by Stoy et al. (2005), who found that the major environmental variables affecting WEC exchanges varied according to the temporal scale.

Figure 13.6 shows the time series of annual precipitation and SWC in the surface soil layer (0–50 cm) during June, July, and August. The interannual variation of SWC lagged that of annual precipitation by one year. Summer precipitation would be stored in the surface soil, which froze from late autumn to the following spring. Thus, the permafrost has a 1-year memory of precipitation and functions as a buffer for the interannual variation of precipitation. Because of this memory function, we could not find clear relationships between evapotranspiration and precipitation, and SWC was the most important factor for E/E_p . The interannual fluctuation of plant activity has also been examined by dendrochronology. Rigling et al. (2001) used tree-ring analyses to

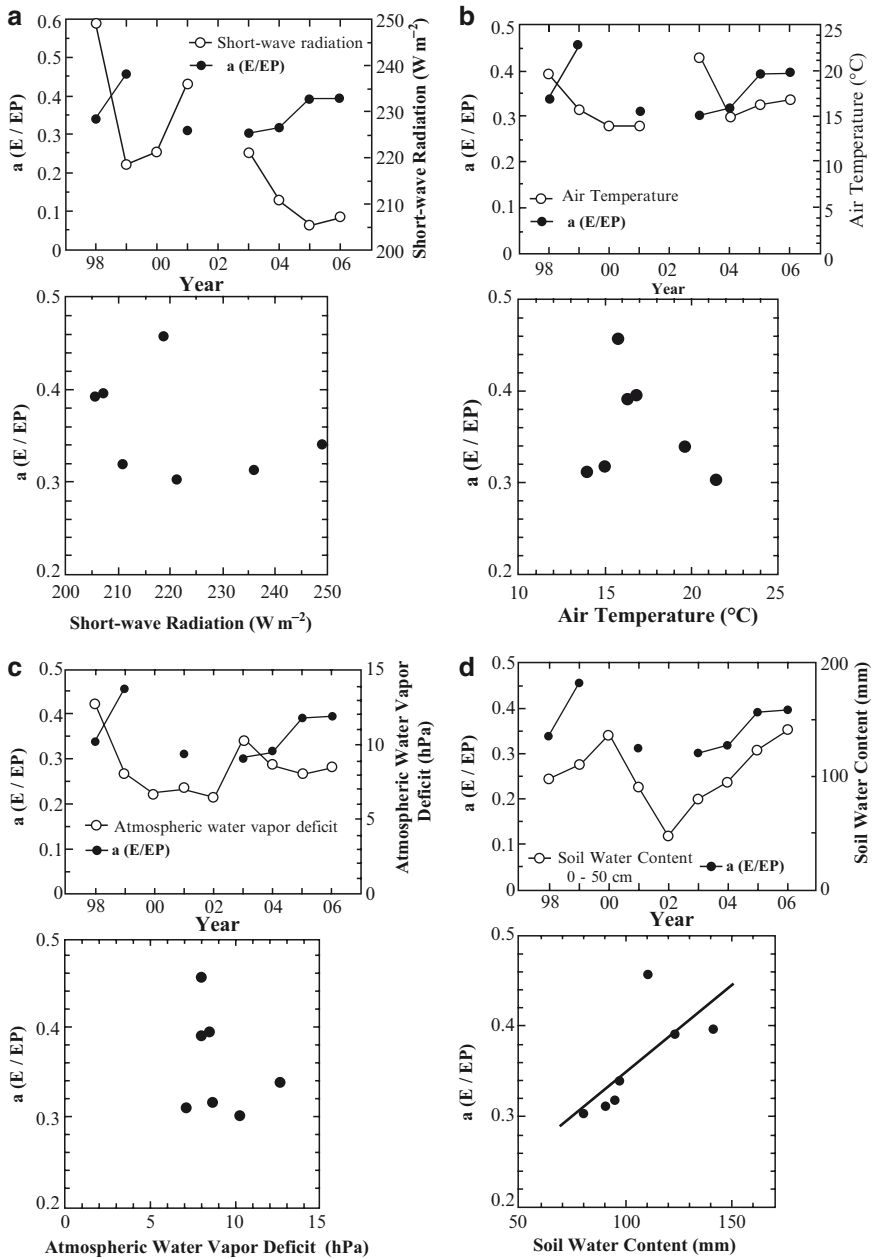


Fig. 13.5 Interannual time series and relationships of environmental variables and evapotranspiration normalized by potential evapotranspiration (E/Ep). (a) Shortwave radiation; (b) air temperature; (c) atmospheric water vapor deficit, and (d) soil water content (SWC). (Ohta et al. 2008)

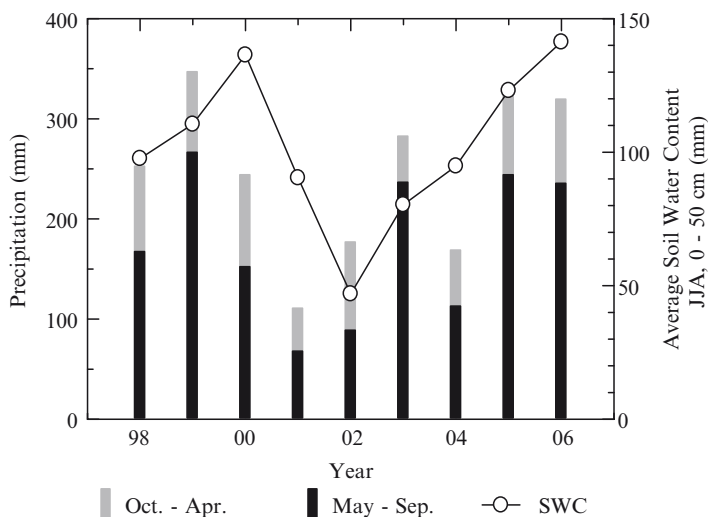


Fig. 13.6 Interannual variation of annual precipitation amount and SWC in the 0–50 cm depth layer. One water year is defined as 365 days from October 1 in the previous year to September 30 in the following year. (Ohta et al. 2008)

examine the relationships between climate and tree growth. They found that tree growth strongly corresponded to the amount of precipitation in the previous summer in Siberian forests. This result further demonstrates that frozen soil has a 1-year memory of precipitation. The 1-year time lag of SWC or plant activity is also found at the stand scale in the region of continuous permafrost in Central Siberia (see Chap. 17, this Vol.).

As mentioned previously, Siberian larch trees uptake permafrost melt water in drought years, and the permafrost reflects interannual variations of precipitation with a 1-year time lag. These two buffers result in the small interannual fluctuation of evapotranspiration, even in a region of relatively low precipitation, and in a significantly different relationship between annual precipitation and evapotranspiration compared with relationships obtained in non-permafrost regions (Zhang et al. 2001). A similar tendency might occur in photosynthetic activities in Siberian larch forests.

13.4.3 Water and Energy Exchange Differences between Non-permafrost and Permafrost Areas of Siberia

While Northeastern and Central Siberia are broadly covered with continuous permafrost, there is less permafrost in Western Siberia (Fig. 1.1, this Vol.). The distribution of the permafrost is also much more extensive in Northeastern Siberia than that in Central Siberia, where its distribution is either discontinuous or sporadic in the southern part of the region. However, few studies have investigated water and energy cycles in those regions. Tchebakova et al. (2002) investigated the water and energy exchanges

in a Scots pine (*Pinus sylvestris*) forest without permafrost at the Zotino site near the eastern boundary of Western Siberia. They examined seasonal variations of energy balance components. The seasonal time series of sensible heat roughly followed that of net all-wave radiation and did not show a clear peak just before the growing season, unlike in the larch forest in Northeastern Siberia. On the other hand, the course of latent heat flux was rather stable during the growing seasons. A similar time series was found at another Scots pine forest near Yakutsk in Northeastern Siberia with permafrost (Hamada et al. 2004). The differences in the seasonal variation of sensible and latent heat fluxes might have resulted from difference in the forest types, i.e., evergreen (pine) versus deciduous (larch), rather than the existence of permafrost.

The amount of evapotranspiration fluctuated from 147 to 196 mm during June, July, August, and September with annual precipitation of 188 to 209 mm in the same period (Tchebakova et al. 2002). The daily evapotranspiration rates were 0.6–2.0 mm d⁻¹. These rates were in the same range as that found in Northeastern Siberia, despite differences in forest types and permafrost existence, but lower than the range for Scandinavian boreal forests (Grelle et al. 1999). The range of fluctuation in evapotranspiration during the growing seasons was wider than that of the precipitation. This result shows clear difference in the interannual variation of evapotranspiration between permafrost (mentioned in Subsect. 13.4.1 and 13.4.2) and non-permafrost regions within the Siberian taiga.

Tchebakova et al. (2002) also concluded that low evapotranspiration of the Scots pine forest in Central Siberia was due to low LAI and low SWC. This conclusion qualitatively agrees with those obtained from the comparative analysis of boreal and temperate forests, as discussed in Subsect. 13.4.4. Surface conductance was controlled by atmospheric water vapor deficit rather than by SWC. This result also agrees with that found for a Northeastern Siberian pine forest (Hamada et al. 2004) where daily surface conductance was regulated by atmospheric water vapor deficit.

Several common features, including low evapotranspiration and low surface conductance, were found to be controlled by atmospheric water vapor deficit in larch forests of Northeastern Siberia and in pine forests of Central Siberia. On the other hand, there were different characteristics of the seasonal course of sensible and latent heat fluxes and the interannual variation of evapotranspiration. However, we cannot conclude which effect was more significant; forest types, existence of permafrost, or both.

13.4.4 Water and Energy Exchange in Different Environments and Climates

13.4.4.1 Seasonal Variations and Magnitudes among Climates

In recent decades, various single-site studies have examined canopy-atmosphere water and energy exchanges in different climates, e.g., tropical forest (Shuttleworth et al. 1984; Malhi et al. 2002; Kumagai et al. 2005), temperate forest (Verma et al. 1986; Kosugi et al. 2006), and boreal forest (Kelliher et al. 1997; Ohta et al. 2001).

These studies have suggested that the latent heat flux or evapotranspiration is generally smaller in boreal forests and that a larger part of available energy is consumed as sensible heat flux in high-latitude areas. However, the major factors characterizing hydrological aspects, especially evapotranspiration, have not yet been clarified.

Matsumoto et al. (2008a) examined differences in water and energy exchange characteristics at five forested sites, including Siberian larch and pine forests, in three climate zones: boreal, cool temperate, and warm temperate. Figure 13.7

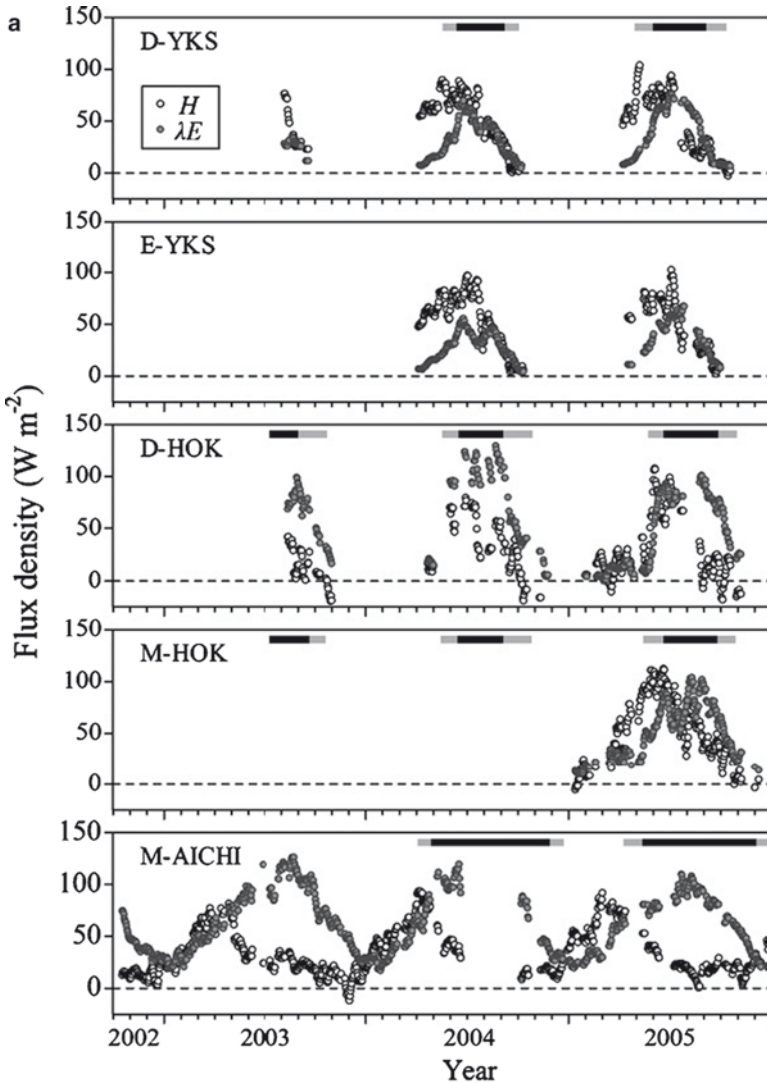


Fig. 13.7 Interannual variation of (a) sensible heat flux (*open circle*) and latent heat flux (*filled circle*) in the Siberian forests (*top two panels*), the cool-temperate forests (*middle two panels*), the warm-temperate forest (*bottom panel*); and (b) the evaporative fraction (Matsumoto et al. 2008a)

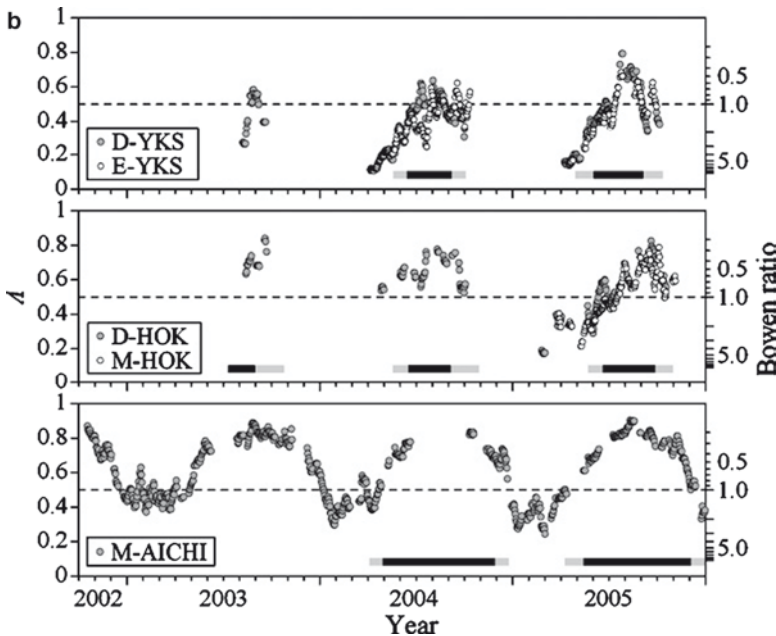


Fig. 13.7 (continued)

shows (a) time series of the latent and sensible heat fluxes and (b) the evaporative fraction Λ , defined by Eq. (13.3) and the Bowen ratio. They found no significant difference in the magnitude and seasonal variation between different forest types located in the same climate zone. However, there were remarkable differences in the exchanges of water and energy among the climate zones. Namely, the growing season, defined as the duration of $\Lambda \geq 0.5$ ($\beta < 1.0$), was longer at sites in the southern area. At northern sites, the magnitude of Λ was lower. The finding of smaller latent heat flux in the northern forests agrees with previous insights obtained from single-site studies. Maximum sensible heat flux occurred earlier than maximum latent heat flux, and time lag of the latent heat flux peak was shorter at northern sites. Regardless of the climate zone, the timing of maximum sensible heat flux coincided with that of leaf emergence of deciduous trees. In the warm-temperate forest, the sensible heat flux had a maximum value in April. Subsequently, the sensible heat flux decreased to only approximately $20\text{--}30\text{ W m}^{-2}$, and the latent heat flux increased to more than 100 W m^{-2} in July and/or August. On the other hand, in the two boreal forests, maximum latent heat flux lagged the peak of sensible heat flux by only 2–4 weeks, and the sensible heat flux remained high (more than 50 W m^{-2}) and somewhat larger than the latent heat flux even during the mid-growing season in two boreal forests. The effect of atmospheric heating was more significant in the boreal forests than in the temperate forests.

13.4.4.2 Factors Controlling the Difference in Water and Energy Exchanges among Climate Zones

In this section, we discuss the major factors causing differences in water and energy exchanges among the climate zones mentioned above. Matsumoto et al. (2008a) compared the atmospheric and land-surface regulations of daily evapotranspiration (E_d ; mm d^{-1}) based on potential evaporation (E_p) and an evapotranspiration coefficient (a). The relationship of the three variables is given as:

$$E_d = aE_p, \quad (13.5)$$

where E_p represents the atmospheric demand for evaporation under a given atmospheric condition and a indicates regulation of evapotranspiration by land surface processes.

Figure 13.8 shows the relationships between the evapotranspiration coefficient and daily evapotranspiration in July and August at each site. Solid lines indicate potential evaporation of 1–6 mm d^{-1} . The values of E_p ranged from 4–4.7 mm d^{-1} in the two Siberian forests and from 4–5.1 mm d^{-1} in the three temperate forests. The atmospheric demand for evaporation in the Siberian forests was quite similar to that in the temperate forests. This result indicates that there was no significant difference in the atmospheric demand for evaporation among the climates and that

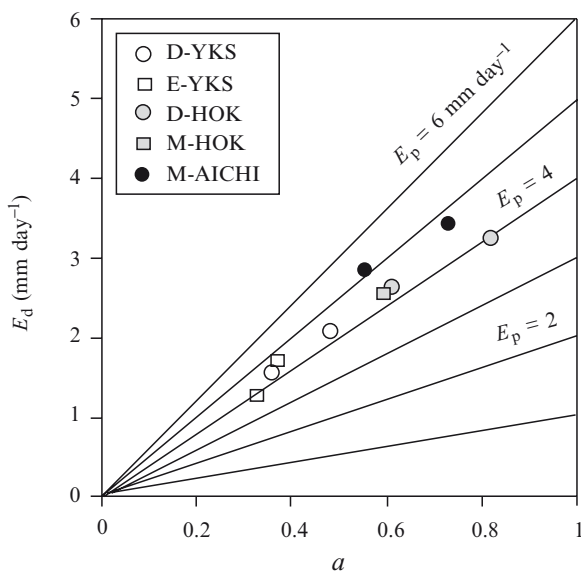


Fig. 13.8 Relationships between the evapotranspiration coefficient a and daily evapotranspiration rates at five forests in Siberia and temperate climate zones. Solid lines indicate potential evaporation for each value presented. (Matsumoto et al. 2008a)

the differences in water and energy exchange did not result from the atmospheric conditions. By contrast, the ranges of a values clearly differed between the boreal and temperate forests, with values of 0.3–0.5 in the Siberian forests and 0.6–0.8 in the temperate forests. These values indicate that evapotranspiration is strongly limited by land surface processes in the Siberian forests and that the evapotranspiration differences among the forests in three climates mainly resulted from the variation in land surface regulation.

Figure 13.9 shows the relationships between the parameter a and the atmospheric vapor pressure deficit (D), SWC (θ), or LAI, and a . Low atmospheric humidity and/or low SWC roughly induced the low evapotranspiration coefficient (Fig. 13.9a, b). However, a was lower in the Siberian forests than in the warm-temperate forest at the same ranges of θ and D . This difference can be explained by the difference in LAI (Fig. 13.9c): the forests with lower LAI had lower a , and there was no clear relationship between the other environmental factors and a .

The results reported by Matsumoto et al. (2008a) also suggested that low evapotranspiration in the Siberian forests might have resulted from land surface processes.

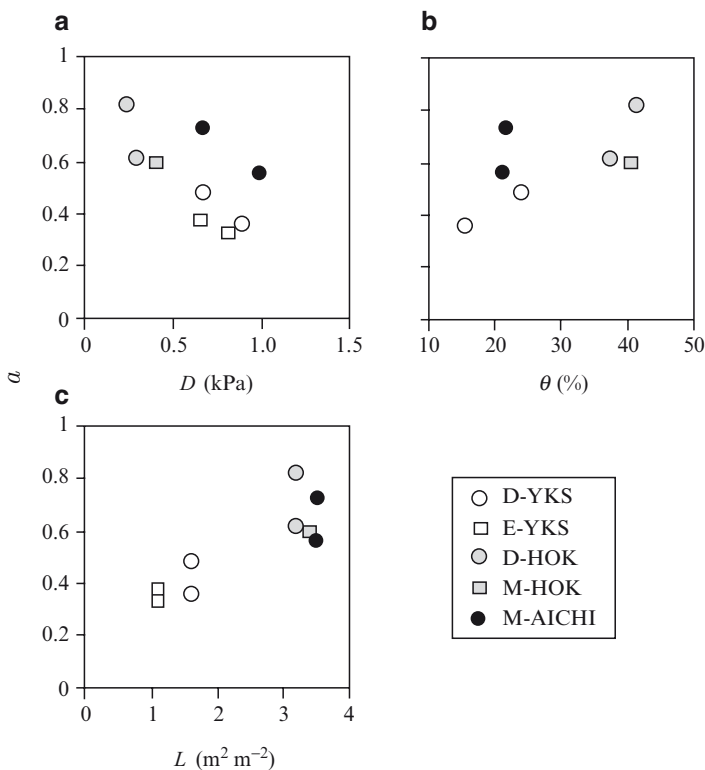


Fig. 13.9 Relationships between the evapotranspiration coefficient a and (a) atmospheric water vapor deficit (D), (b) SWC (θ), and (c) leaf area index (LAI). Here, LAI is indicated as L according to the original figure (Matsumoto et al. 2008a)

In particular, atmospheric and soil moisture conditions and leaf mass (or LAI) appeared to strongly affect the differences in evapotranspiration among three climates considered. Lower leaf mass has a significant effect not only on evapotranspiration but also on net carbon exchange (see Chap. 10, this Vol.).

13.5 Evaluation of Hydrological Aspects in Northeastern Siberia

Previous sections have discussed hydrological aspects at the stand scale. However, regional characteristics cannot be understood only by stand-scale analyses. Here, we examine results obtained from the LSM and DRM analyses to understand regional-scale hydrological aspects over Central and Northeastern Siberia.

The results of the LSM (Park et al. 2008) and the DRM (Hatta et al. 2009) agreed well with the annual water balance and hydrographs at four major hydrological observation stations located along three tributaries and the mouth of the Lena River (results not shown). Table 13.2 shows the precipitation and evapotranspiration obtained by four studies. The evapotranspiration estimated by Park et al. (2008) agreed well with the results of previous studies, suggesting that Northeastern Siberia has unique surface conductance responses and supporting Matsumoto et al.'s (2008b) hypothesis that surface conductance responses against environmental variables are potentially similar throughout Siberia, despite wide variation in tree species and forest types. Figure 13.10 shows fields of mean net all-wave radiation, latent and sensible heat fluxes, and ground heat flow over Central and Northeastern Siberia during the period of 1986–2004 (Park et al. 2008). According to their results, energy consumption into the latent heat flux was twice as large ($\sim 25 \text{ W m}^{-2}$) in the boreal forests than in the tundra region. Tundra with low LAI might have low surface conductance due to rapid desiccation (Beringer et al. 2005). In addition, higher transpiration in the forested area would result from higher LAI compared with the tundra region. A dominant component of the energy balance in the boreal forest region was latent heat flux ($\sim 50 \text{ W m}^{-2}$), while a dominant component in the tundra region was sensible heat flux ($\sim 35 \text{ W m}^{-2}$). However, the magnitude of sensible heat flux was smaller in the tundra region than in the forest region

Table 13.2 Estimated annual precipitation and evapotranspiration in the Lena River basin

Precipitation (mm y^{-1})	Evapotranspiration (mm y^{-1})	Source
368.9	195.9	Park et al. (2008)
403.0	224.0 ^a	Serreze et al. (2003)
	182.0 ^b	Serreze et al. (2003)
350.0	160.0	Fukutomi et al. (2003)

The values of a and b were calculated by atmospheric water budget and water budget analyses using measured runoff datasets, respectively

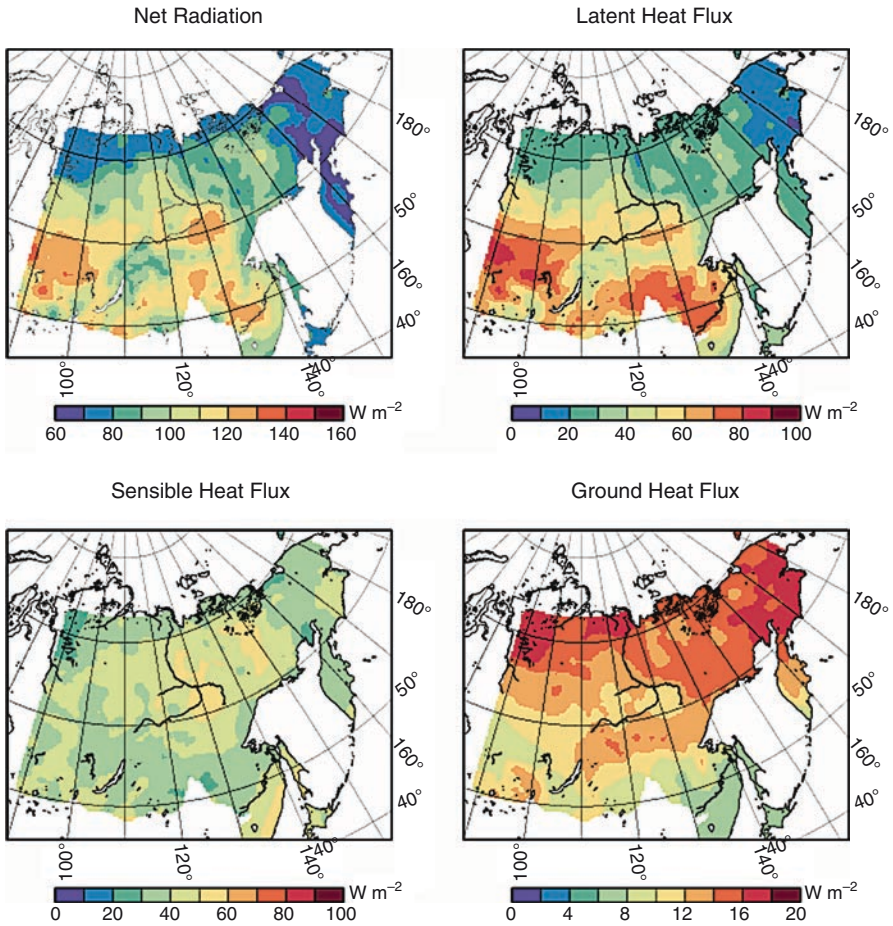


Fig. 13.10 Spatial mean distribution of net all-wave radiation, latent heat flux, sensible heat flux, and ground heat flow over Central and Northeastern Siberia during a period 1986–2004, estimated by a land surface model (Park et al. 2008) (see Color Plates)

($\sim 45 \text{ W m}^{-2}$) because available energy, $R_n - G$, was much higher in the forest region ($\sim 110 \text{ W m}^{-2}$) than in the tundra region ($\sim 70 \text{ W m}^{-2}$) due to low albedo. The contrast in the energy balance between boreal forest and tundra occurred in the magnitude of net all-wave radiation due to albedo and in latent heat flux due to leaf mass.

Park et al. (2008) also examined the interannual variation of latent heat flux and its controlling factor over Northeastern Siberia. They found a significantly positive relationship between annual mean temperature during May–August and annual evapotranspiration over Northeastern Siberia. This implies that summer air temperature strongly controlled annual evapotranspiration, in agreement with results obtained by Suzuki et al. (2006) using the National Oceanic and Atmospheric Administration (NOAA)/advanced very high resolution radiometer dataset.

These results contradict findings obtained at stand scales, mentioned in Subject 13.4.2 (Rigling et al. 2001; Ohta et al. 2008). Issues of spatial scaling may explain the difference. The model result has an advantage in that it spatially extends the representation, but at the cost of accuracy, for variables such as SWC. The opposite holds true for single-stand studies.

In addition, Park et al. (2008) showed a negative relationship between annual evapotranspiration and precipitation. Fukutomi et al. (2003) found an inverse relationship between these two variables in the Lena River basin, but they found no inverse relationship in the Ob and Yenisei watersheds. The Ob and Yenisei watersheds have much less permafrost coverage than the Lena watershed. This interesting contrast among the Siberian basins might result from permafrost conditions.

13.6 Conclusions

This section has focused on hydrological aspects of Northeastern Siberian forests, which strongly affect ecological characteristics. The seasonal and interannual variations of water and energy exchanges between forests and the atmosphere were discussed based on observations using an eddy covariance technique. The characteristics of Siberian forests were compared with those of forests located in other climate zones of East Asia. Finally, hydrological features over Central and Northeastern Siberia were examined based on results of the LSM analysis. Aspects of water and energy exchanges in Siberian forests are summarized as follows:

- The timing of latent heat flux increase and larch leaf emergence coincided. At the same time, sensible heat flux dropped, although net all-wave radiation continued to increase. However, the magnitude of latent heat flux was almost the same as that of sensible heat flux even in the mid-growing season. Snow melt did not significantly affect the water and energy exchanges between the forest and atmosphere, even in the mid-growing season, although snow melt was significant from a hydrological perspective.
- Annual evapotranspiration did not fluctuate widely, although annual precipitation showed a wide range. Annual evaporation rates were usually sensitive to changes in annual precipitation in non-permafrost regions, but the opposite was true in Northeastern Siberia. This difference in Northeastern Siberia was attributable to interactions between permafrost and vegetation. Interannual variation of surface soil water lagged that of annual precipitation by 1 year. In addition, trees can use permafrost melt water during drought years. Permafrost thus acts as a buffer against fluctuations in annual precipitation.
- The low evapotranspiration in the Siberian forests resulted from strong land surface regulation compared with that in temperate forests. Low SWC, high atmospheric water vapor deficit, and less leaf mass led to the low evapotranspiration rates in Northeastern Siberia.

- Annual evapotranspiration was strongly affected by soil moisture content. This characteristic resulted from the 1-year lag of soil moisture mentioned above. However, contradictions were found between the stand-scale analyses and watershed- or regional-scale investigations using an LSM or satellite data. Such discrepancies might have been due to issues with scaling.
- At a watershed scale, hydrometeorological aspects in the Siberian taiga region differed significantly from those in the tundra region. The magnitude of sensible heat flux in the tundra region was similar to that of latent heat flux in the Siberian taiga region, although the dominant turbulent flux component was sensible heat flux in the tundra region. However, the sensible heat flux was smaller in the tundra region than in the taiga region because of the high albedo of the former vegetation. In addition, a negative relationship between annual precipitation and evapotranspiration was only found in the Lena River basin. This interesting phenomenon might have resulted from the extensive permafrost in the Lena River basin.

As discussed above, hydrological aspects in Northeastern Siberia are strongly affected by interactions among the atmosphere, vegetation, and soil and are particularly affected by the existence of permafrost. Permafrost appears to function as a strong buffer for the wide fluctuations of annual precipitation in the region; by this permafrost buffer, the Siberian taiga forest can maintain stable evapotranspiration and probably photosynthesis. Several recent studies (e.g., Peterson et al. 2002; Yang et al. 2002; Serreze et al. 2003; McClelland et al. 2004) have reported increases in river runoff both in North America and Siberia. Runoff can be estimated as precipitation minus evapotranspiration; thus, understanding evapotranspiration processes will contribute to understanding of not only ecological features but also hydrological properties. Siberian forests also affect the spatial distribution of precipitation. Yoon and Cheng (2006) suggested that evapotranspiration from Siberian forests might maintain the rain belt along the high latitudes.

Woo et al. (2008) emphasized the importance of surface–subsurface processes in the hydrological processes of the permafrost region. Vegetation should also be considered in the examinations of the water cycles in this region. Understanding the interactions of hydrological and ecological factors will contribute to understandings of both of these scientific fields in the permafrost regions.

References

- Baldocchi D, Falge E, Wilson K (2001) Asepectral analysis of biosphere-atmosphere trace gas flux densities and meteorological variables across hour to multi-year time scales. *Agr For Meteorol* 107:1–27
- Barr AG, Morgenstern K, Black TA, McCaughey JH, Nesic Z (2006) Surface energy closure by the eddy-covariance method above three boreal stands and implication for the measurement of the CO₂ flux. *Agr For Meteorol* 140:322–337
- Beringer J, Chapin FS III, Thompson CC, McGuire AD (2005) Surface energy exchanges along a tundra-forest transition and feedbacks to climate. *Agr For Meteorol* 131:143–161

- Bonan GB, Pollard D, Thompson SL (1992) Effects of boreal forest vegetation on global climate. *Nature* 359:716–718
- Bonan GB, Chapin FS III, Thompson SL (1995) Boreal forest and tundra ecosystems as components of the climate system. *Clim Change* 29:145–167
- Dolman AJ, Maximov TC, Moors EJ, Maximov AP, Elbers JA, Kononov AV, Waterloo MJ, van der Molen MK (2004) Net ecosystem exchange of carbon dioxide and water of far eastern Siberian Larch (*Larix cajanderii*) on permafrost. *Biogeosciences* 1:133–146
- Fukutomi Y, Igarashi H, Masuda K, Yasunari T (2003) Interannual variability of summer water balance components in three major river basins of northern Eurasia. *J Hydrometeorol* 4:283–296
- Grelle A, Lindroth A, Molder M (1999) Seasonal variation of boreal forest surface conductance and evaporation. *Agr For Meteorol* 98–99:563–578
- Halldin S, Gottschalk L, Griend AA, Gryning S-E, Heikinheimo M, Hogstrom U, Jochum A, Lundin L-C (1998) NOPEX—a Northern hemisphere climate processes land surface. *J Hydrol* 212–213:172–187
- Hamada S, Ohta T, Hiyam T, Kuwada T, Takahashi A, Maximov TC (2004) Hydrometeorological behaviour of pine and larch forests in eastern Siberia. *Hydrol Proc* 18:23–29
- Hatta S, Hayakawa H, Park H, Yamazaki T, Yamamoto K, Ohta T (2009) Long term runoff analysis of the Lena river basin using a distributed hydrological model. *J Jpn Soc Hydrol Water Resour* 22:177–187 in Japanese with English abstract
- Katul G, Lai C-T, Schafer K, Vidakovic B, Albertson J, Ellsworth D, Oren R (2001) Multiscale analysis of vegetation surface fluxes: from seconds to years. *Adv Water Resour* 24:1119–1132
- Kelliher FM, Hollinger DY, Schulze E-D, Vygodskaya NN, Byers JN, Hunt JE, McSeveny TM, Milukova I, Sogatchev A, Varlargin A, Ziegler W, Arneth A, Bauer G (1997) Evaporation from an eastern Siberian larch forest. *Agric For Meteorol* 85:135–147
- Kelliher FM, Lloyd J, Arneth A, Byers JN, McSeveny TM, Milukova I, Grigoriev S, Panfyorov M, Sogatchev A, Varlargin A, Ziegler W, Bauer G, Schulze E-D (1998) Evaporation from a central Siberian pine forest. *J Hydrol* 205:279–296
- Kosugi Y, Takanashi S, Matsuo N, Tanaka K, Tanaka H (2006) Impact of leaf physiology on gas exchange in a Japanese evergreen broad-leaved forest. *Agr For Meteorol* 139:182–199
- Kosugi Y, Takanashi S, Tanaka H, Ohkubo S, Tani M, Yano M, Katayama T (2007) Evapotranspiration over a Japanese cypress forest. I: Eddy covariance fluxes and surface conductance characteristics for 3 years. *J Hydrol* 337:269–283
- Kumagai T, Saitoh TM, Sato Y, Takahashi H, Manfroi OJ, Morooka T, Kuraji K, Suzuki M, Yasunari T, Komatsu H (2005) Annual water balance and seasonality of evapotranspiration in a Bornean tropical rainforest. *Agr For Meteorol* 128:81–92
- Lawford RG, Stewart R, Roads J, Isemer H-J, Manton M, Marengo J, Yasunari T, Benedict S, Koike T, Williams S (2004) Advancing global and continental-scale hydrometeorology. *Bull Am Met Soc* 85:1917–1930
- Lu M, Koike T, Hayakawa N (1989) A rainfall-runoff model using distributed data of radar rain and altitude. *Proceedings of Jpn Soc Civil Engin* 411 II-12:135–142 (in Japanese with English abstract)
- Malhi Y, Pegoraro E, Nobre AD, Pereira MGP, Grace J, Culf AD (2002) Energy and water dynamics of a central Amazonian rain forest. *J Geophys Res* 107(D20):LBA 45.1–LBA 45.17. doi:10.1029/2001JD000623
- Matsumoto K, Ohta T, Nakai T, Kuwada T, Daikoku K, Iida S, Yabuki H, Kononov AV, van der Molen MK, Kodama Y, Maximov TC, Dolman AJ, Hattori S (2008a) Energy consumption and evapotranspiration at several boreal and temperate forests in the Far East. *Agr For Meteorol* . doi:10.1016/j.agrformet.2008.09.008
- Matsumoto K, Ohta T, Nakai T, Kuwada T, Daikoku K, Iida S, Yabuki H, Kononov AV, van der Molen MK, Kodama Y, Maximov TC, Dolman AJ, Hattori S (2008b) Responses of surface conductance to forest environments in the Far East. *Agr For Meteorol* . doi:10.1016/j.agrformet.2008.09.009

- McClelland JW, Holmes RM, Peterson BJ (2004) Increasing river discharge in the Eurasian Arctic: consideration of dams, permafrost thaw, and fires as potential agents of change. *J Geophys Res* 109:D18102. doi:[10.1029/2004JD004583](https://doi.org/10.1029/2004JD004583)
- Moors EJ, Dolman H, Maximov TC, Jans W, Maximov A, Kononov A, Kruijt B, Nabuurs GJ (2006) Results of the PINMATRA project 2002-2004. In: Proceedings of Second International Workshop on C/H₂O/Energy Balance and Climate over Boreal Regions with Special Emphasis on Eastern Siberia. pp 3–6 <http://www.agr.nagoya-u.ac.jp/~wecnof/>
- Ohta T, Hiyama T, Tanaka H, Kuwada T, Maximov TC, Ohata T, Fukushima Y (2001) Seasonal variation in the energy and water exchanges above and below a larch forest in eastern Siberia. *Hydrol Proc* 15:1459–1476
- Ohta T, Maximov TC, Dolman AJ, Nakai T, van der Molen MK, Kononov AV, Maximov AP, Hiyama T, Iijima Y, Moors EJ, Tanaka H, Toba T, Yabuki H (2008) Interannual variation of water balance and summer evapotranspiration in an eastern Siberian larch forest over a 7-year period (1998–2006). *Agr For Meteorol*. doi:[10.1016/j.agrformet.2008.04.012](https://doi.org/10.1016/j.agrformet.2008.04.012)
- Park H, Yamazaki T, Yamamoto K, Ohta T (2008) Tempo-spatial characteristics of energy budget and evapotranspiration in the eastern Siberia. *Agr For Meteorol*. doi:[10.1016/j.agrformet.2008.06.018](https://doi.org/10.1016/j.agrformet.2008.06.018)
- Peterson BJ, Holmes RM, McClelland JW, Vorosmarty CJ, Lammers RB, Shiklomanov AI, Shiklomanov IA, Rahmstorf S (2002) Increasing river discharge to the Arctic Ocean. *Science* 298:2171. doi:[10.1126/science.1077445](https://doi.org/10.1126/science.1077445)
- Rigling A, Waldner PO, Forster T, Brake OU, Pouttu A (2001) Ecological interpretation of tree-ring width and intraannual density fluctuations in *Pinus sylvestris* on dry sites in the central Alps and Siberia. *Can J For Res* 31:18–31
- Schulze E-D, Vygodskaya NN, Tchepakova NM, Czimeczik CI, Kozlov DN, Lloyd J, Mollicone D, Parfenova E, Sidorov KN, Varlagin AV, Wirth C (2002) The Eurosiberian transect: an introduction to the experimental region. *Tellus* 54B:421–428
- Sellers P, Hall F, Margolis H, Kelly B, Baldocchi D, Hartog GD, Cihlar J, Ryan MG, Goodison B, Crill P, Ranson KJ, Lettenmaier D, Wickland DE (1995) The boreal ecosystem—atmosphere study (BOREAS): an overview and early results from the 1994 field year. *Bull Am Meteorol Soc* 76:1549–1577
- Serreze MC, Bromwich DH, Clark MP, Etringer AJ, Zhang TZ, Lammers R (2003) Large-scale hydro-climatology of the terrestrial Arctic drainage system. *J Geophys Res* 108(D2):8160. doi:[10.1029/2001JD000919](https://doi.org/10.1029/2001JD000919)
- Shuttleworth WJ, Gash JHC, Lloyd CR, Moore CJ (1984) Eddy correlation measurements of energy partition for Amazonia. *Q J Roy Meteorol Soc* 110:1143–1162
- Stoy PC, Katul GG, Siqueira MBS, Juang J-Y, McCarthy HR, Kim H-S, Oishi AC, Oren R (2005) Variability in net ecosystem exchange from hourly to inter-annual time scales at adjacent pine and hardwood forests: a wavelet analysis. *Tree Physiol* 25:887–902
- Sugimoto A, Yanagisawa N, Naito D, Fujita N, Maximov TC (2002) Importance of permafrost as a source of water for plants in east Siberian taiga. *Ecol Res* 17:493–503
- Suzuki R, Xu J, Motoya K (2006) Global analyses of satellite derived vegetation index related to climatological wetness and warmth. *Int J Climatol* 26:425–438
- Tchepakova NM, Kolle O, Zolotoukhine D, Arneth A, Styles JM, Vygodskaya NN, Schulze E-D, Shibistova O, Lloyd J (2002) Inter-annual and seasonal variations of energy and water vapour fluxes above *Pinus sylvestris* forest in the Siberian middle taiga. *Tellus* 54B:537–551
- Verma SB, Baldocchi DD, Anderson DE, Matt DR, Clement RJ (1986) Eddy fluxes of CO₂, water vapor, and sensible heat over a deciduous forest. *Boundary-Layer Meteorol* 36:71–91
- Woo M-K, Kane DL, Carey SK, Yang D (2008) Progress in permafrost hydrology in the new millennium. *Permafrost Periglacial Proc* 19:237–254
- Yamazaki T (2001) A one-dimensional land surface model adaptable to intensely cold regions and its application in eastern Siberia. *J Meteorol Soc Jpn* 79:1107–1118
- Yang D, Kane DL, Hinzman LD, Zhang X, Zhang T (2002) Siberian Lena river hydrologic regime and recent change. *J Geophys Res* 107(D23):4694. doi:[10.1029/2002.JD002543](https://doi.org/10.1029/2002.JD002543)

- Yasunari T (2007) Role of land atmosphere interaction on Asian monsoon climate. *J Meteorol Soc Jpn* 85B:55–75
- Yoon J-H, Cheng T-C (2006) Maintenance of the boreal forest rainbelts during northern summer. *J Climate* 19:1437–1449
- Zhang L, Dawes WR, Walker GR (2001) Response of mean evapotranspiration to vegetation change at catchment scale. *Water Resour Res* 37:701–708