Evaporation from land surface in high latitude areas: a review of methods and study results*

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Abstract Evaporation (ET) from land surfaces in high latitudes is examined on a circumpolar perspective based upon the study results obtained in various environments, from boreal forest (taiga) to the high Arctic desert. Direct and indirect methods of evaporation measurement are reviewed, as well as numerous computational techniques. We have focused upon methods conveniently adopted for calculating evaporation when detailed information on meteorological conditions within the surface boundary layer is not available. These methods range from complicated ones, such as eddy correlation, energy balance and Penman equations, to empirical relationships between ET and incoming solar radiation. Great attention was paid to the principles of each method, especially those developed in Russia as they differ from most of the methods utilized internationally. For example, the Budyko–Zubenok empirical scheme is based upon the principle of potential evaporation, which is affected by soil moisture (SM). This relationship between ET and SM, expressed in terms of the field capacity, has been found to be non-linear; a complication that is not typically accounted for in traditional approaches. This paper also contains a brief review of a number of evaporation case studies including Alaska (USA), north-western Russia and Siberian taiga, Yukon basin (Canada), mountainous forest on Hokkaido Island (Japan), Canadian Arctic and glacierized basins of Greenland.

Keywords Case studies; computational methods; evapotranspiration; field measurements; high latitudes; synthesis

Introduction

It is a significant challenge to obtain good flux estimates of water as a liquid or solid (ice), from the ground surface to the atmosphere as vapor in high latitude basins because of the cold environment. The upward migration of water vapor back to the atmosphere is a significant component of the hydrologic cycle that we should attempt to measure more accurately at the watershed scale. Also, sublimation from the winter snow cover is a component of the vertical vapor flux that is very poorly quantified at the present time.

Both evaporation and evapotranspiration (ET) from land surface can be determined by either direct or indirect measurements that include various computational methods. The only direct methods in hydrologic research to measure ET rates are the traditional methods of water balance or lysimetric studies. Lysimeter results represent a very small area, while water balance studies provide results averaged over a larger area (catchment scale) and give no insight about spatial variability. Also, when the water balance equation is used, all of the measurement errors present in the other equation terms, as well as those terms that are

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ignored, are combined in the estimate of ET. Some 39 water balance studies which included an ET determination were presented at a special Northern Research Basins workshop (Kane and Yang 2004).

Although used in early studies in Russia, lysimeter measurements under Arctic conditions are seldom performed because of the difficulties associated with frozen ground. More typically a range of ET results are obtained, in particular, by clarifying empirical factors impacting estimated evaporation rates (Marsh *et al.* 1981; Roulet and Woo 1986; Mendez *et al.* 1998). It is clear that, more frequently, the present determination of ET from land surfaces in Arctic and sub-Arctic environments are based on one of a variety of calculation methods.

The most applicable outside of Russia are the methods developed by Penman (1956) and Priestley and Taylor (1972). These methods are not used nor accepted widely in Russia. Therefore, we would like to introduce and, if possible, to test them in comparison with the original Russian methods and observational results. The long-term evaporation studies conducted in Valday, Russia present an excellent data source to evaluate and compare the empirical techniques developed in various regions.

In the remainder of this paper, we present a short discussion on the numerous methods used, with a special emphasis on the Russian methods. This is followed by several case studies from around the northern circumpolar region. Finally, a summary of regional evaporation rates is presented.

Methods

Direct lysimeter observations

Evaporation from vegetated surfaces, bare soil and open water, were measured at Valday for over 40 years going back to the early 1950s. There have been three types of devices used: a set of weighing lysimeters with 0.3 m^2 surface area and 0.6 m depth (up to 16 units for different crops), a large floating lysimeter (5 m^2 in area) for measurement of hourly rates and four deep (1.8 m) lysimeters (0.3 m^2 in area) with controllable ground water table to compensate for evaporative losses. Evaporation rates from open water surfaces were measured in tanks with pan areas of 0.3 m^2 and 20 m^2, both at 0.6 m depth. The observational data for long-term periods are thoroughly examined in a case study below.

Lysimetric measurements in Arctic conditions are seldom performed because of the difficulty in maintaining water tanks and pipes in frozen ground. Numerous studies have demonstrated that soil moisture content availability greatly affects evaporation (Marsh *et al.* 1981; Kane *et al.* 1990; Mendez *et al.* 1998). When soil moisture (SM) $> 30\%$ (volumetric percent), evaporation E proceeds at a rate near its potential. As SM decreases, E decreases abruptly; when SM decreases $< 20\%$, E decreases more slowly.

Eddy correlation

Evaporation into the atmosphere is due to turbulent (eddy) flux of moisture content (specific humidity) which can be defined through the following equation:

$$Q_e = \rho L w' q' \qquad (1)$$

where ρ is air density (kg/m^3), L is the latent heat of vaporization ($L = 2.51$ MJ/kg), and w' and q' are fluctuations of vertical wind velocity and the specific air humidity. The prime above signifies in-time averaging of fluctuations $w'q'$.

To measure the eddy correlation $w'q'$ within the atmospheric surface boundary layer requires special, low inertia techniques such as ultrasonic wind velocity sensors and infrared hygrometers. In Russia, an early study with the eddy correlation method was made within and above forest canopies using very small cup type anemometers and low inertia thermo-resistors

(Dubov *et al.* 1978). A classic variant (Kolmogorov 1942) of the method uses the interval of the fluctuation spectra (which can be specified by the so-called " − 2/3 law"). Using average values of fluctuations within that spectral interval allow for a less sensitive measurement technique (Kaimal *et al.* 1972). Nevertheless, the eddy correlation method has never been operationally used for evapotranspiration measurements.

Energy balance: Bowen ratio method

This method is an indirect measurement based upon the following energy balance equation:

$$Q_e = Q_{net} + Q_h + Q_c \tag{2}$$

where Q_e = energy utilized for evaporation or combined ET away from surface (W/m^2), Q_{net} is the net radiation (W/m^2), Q_h is the sensible (convective) heat flux between atmosphere and surface (W/m^2) and Q_c is heat flux into the soil (W/m^2).

The components of Equation (2) can be determined independently pending the determination of the corresponding turbulent exchange coefficient to be used. Another approach is based upon the Bowen ratio, $\beta = Q_h/Q_e$, i.e. the ratio between sensible and latent energy fluxes. It is practical (if the turbulent exchange coefficients for Q_h and Q_e are assumed to be equal) to assume that the ratio $\beta = 0.66\Delta T/\Delta e$, where ΔT and Δe are gradients of temperature and vapor pressure. The exchange coefficients can, through the use of the gradients, be substituted and Equation (2) may be re-written as follows:

$$-Q_e = \frac{(Q_{net} + Q_c)}{(1+\beta)}. \tag{3}$$

This relationship has been examined in many studies carried out on a wide range of surfaces and climatic conditions in high latitudes: from Finland (Tattari 1994) to Eastern Siberian taiga (Vasilenko 2004) and extreme Arctic tundra (Stewart and Rouse 1976; Ishii *et al.* 2004a). All the experiments based on the Bowen ratio method require special attention to the measurement accuracy because low gradients can impart significant errors in determining the energy fluxes. It should also be remembered that averaging the highly variable solar radiation flux may create problems when relating it to other fluxes. It was clearly demonstrated that the most important source of errors derives from the variability of solar radiation due to rapidly changing cloudiness. Individual data errors can be about 30%. Hence, the measurements of radiative fluxes must be integrated over time before they may be used in Equation (2). Another problem appears when small gradients within the surface layer occur (this is a common occurrence for high latitudes). Ohmura (1982) has given criteria for the applicability of the Bowen ratio method.

Profile method

Generally, profile methods imply the detection of the turbulent diffusion of water vapor in the atmospheric surface layer: as such, they may be assumed to be an indirect measurement method. The basic equation is as follows:

$$E = -C_E u_2 F(Ri)(e_2 - e_{0.5}) \tag{4}$$

where E is the evaporation (cm per time unit), C_E is turbulent transfer coefficient for vapor in stable conditions (cm^{-1}) which depends mainly on the aerodynamic roughness of the surface, u_2 is wind speed (m/s) at 2 m level, e_2 and $e_{0.5}$ are vapor pressures (mbar) at 2 m and 0.5 m respectively, and Ri is the Richardson number which is used to define stability (an alternative is the Monin–Obukhov parameter). One expression for the Richardson number is

$$Ri = \frac{-9.8(T_{0.5} - T_2)}{u_2 T_2} \tag{5}$$

where $T_{0.5}$ and T_2 are air temperatures (°C) at 0.5 and 2 m levels. Since the function $F(Ri)$ can only be determined empirically, the approach is often called the semi-empirical theory of turbulent transfer.

The problem is which of the existing variety of $F(Ri)$ functions (Figure 1) is best for the study conditions. As was found by Shutov (2000), about 94% of available data on the Richardson number ranged from -0.04 to 0.1. It is also true that all expressions of the function of $F(Ri)$ give very similar results; they deviate from each other by less than 10%. We offer the following combination of expressions, which have been used in earlier studies (Bush 1976; Thom 1975) and assume $Ri > 0.4$ to avoid singularity:

$$F(Ri) = \begin{cases} (1 - 3\,Ri)^2 & Ri < 0 \\ (1 - 5Ri)^2 & 0 < Ri < 0.4 \\ 0 & Ri > 0.4 \end{cases} \tag{6}$$

These relationships were validated experimentally (Shutov 1998) and can be applied to the empirical determination of evapotranspiration. It must be noted that this combination has been successfully used to evaluate hourly snow melt intensities (Shutov 1991).

Penman equation and its variances

Lake evaporation is often calculated using Penman's combination of energy balance and aerodynamic model, defined as follows (Penman 1956):

$$E = \frac{E_R s + E_A \gamma}{s + \gamma} \tag{7}$$

where $s = 4098e_s/(T + 273)$ and is the slope of the specific humidity and air temperature curve, $\gamma = C_P P_A/0.622L$ is the psychrometric constant, $C_P = 1006$ J/kg°C, $L = 2.51$ MJ/kg defined above, and e_s is saturated vapor pressure at air temperature T (°C). Thus, $E_R = (Q_{\text{net}} - Q_c)/L$ (J/cm^2) describes the "purely" energy-balance component of the evaporation flux. Therefore, $E_A = (e_s - e_2)F(u_2)/\log(z_2/z_0)^2$ is the aerodynamic (advective) component of the entire flux (J/cm^2).

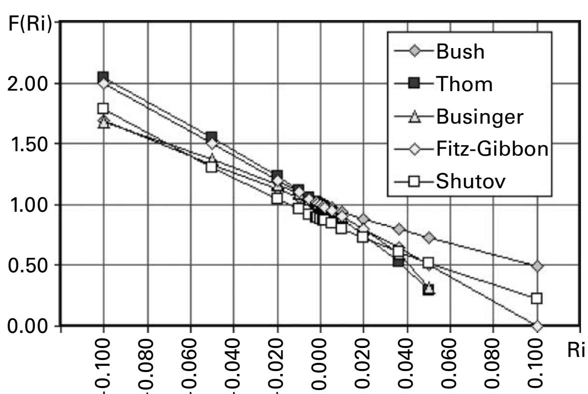

Figure 1 Stability functions $F(Ri)$ offered by several researchers. The most frequent values (more than 90% of all observations) of Ri numbers at Valday range from -0.020 to 0.080

One of the many variants, the Priestley–Taylor Equation (see below) can be obtained by assuming $E = \alpha E_R$, where α is an empirical parameter (defined below). Taking only E_A into account results in the simple Dalton Equation (Brutsaert 1982). Morton (1975) has offered a regional empirical relationship for the stability function $F(u_2)$. This approach to approximate E_A was used for Canadian conditions and seems to be convenient (and often sufficient) for evaporation from lakes. Accounting for the complex effect of the aerodynamic resistance of plants, Monteith (1973) has developed a model for evapotranspiration from vegetated surfaces which is an extension of the original Penman equation and is known as the Penman–Monteith method.

A simplified approach to the Penman method can be obtained by estimating the parameters and relative role of the components in Equation (7). When the air temperature is 0°C, $\gamma \approx 0.66$, $s/(s+\gamma) = 0.4$ and $\gamma/(s+\gamma) = 0.6$. Thus, $E \approx 0.161(Q_{\mathrm{net}} - Q_{\mathrm{c}}) + 0.0096(e_{\mathrm{s}}^* - e_2)u_2$ (in mm/h) when $z_0 = 0.01$ m, which corresponds to a grassland surface.

The advective component appears to be relatively small, especially during conditions of low gradients of vapor pressure that are common for the high latitudes. Indeed, when $u_2 = 3$ m/s and $(e_{\mathrm{s}} - e_2) = 1$ mbar, the evaporation rate is 0.028 mm/h; this will produce less than 0.5 mm/d of evaporation because the nocturnal gradients fall to zero or even become negative. So far, we have not paid great attention to atmospheric instability since most of the necessary input data to consider this are not typically available, except during some experimental field campaigns.

In forested areas, canopy resistance can play an important role as implied in the Penman–Monteith procedure where large variations in evaporation rates occur due to changes in processes that influence the surface energy balance (Baldocchi *et al.* 2000). Canopy resistance greatly influences the rate of transpiration among various tree species. For instance, Higuchi *et al*. (2001) showed that the evaporation flux increased rapidly when larch stands of Eastern Siberia began to foliate. Thus, plant physiology strongly affects the seasonal variation of energy budget and water balance.

Calculating evapotranspiration: Priestley–Taylor method

In the case of a permanently wet surface, the advective component in Equation (6) is assumed negligible and the calculation can be made using the Priestley and Taylor (1972) equation:

$$E = \frac{\alpha s(Q_{\mathrm{net}} - Q_{\mathrm{c}})}{L(s+\gamma)} \tag{8}$$

where Q_{net} is net radiation (W/m^2), Q_{c} is energy flux into the ground (W/m^2), α is an empirical parameter that relates actual to equilibrium evaporation ($\alpha = 1$) and needs to be calibrated to a particular surface type; the other terms are the same as defined earlier for Equation (7). This equation is chosen by many investigators due to the minimal amount of input data required.

One question pervades every application of the Priestley–Taylor method: what value should be selected for the α factor? Priestley and Taylor (1972) assumed it was constant, $\alpha = 1.26$, but other researchers (Davies and Allen 1973) believed the factor ranged from 1.06 to 1.29. The factor α in high latitude areas is often assumed to be dependent on the air temperatures. It is suggested (de Bruin and Keijman 1979) that there are seasonal and even diurnal variations in the value of α. It was found that α is linked with the Bowen ratio. Obviously, the factor α depends on moisture from the evaporated surface, and this problem is addressed, particularly as related to the spatial and temporal variability of evapotranspiration. Kane *et al*. (1990) found that $\alpha =$ 0.95 for upland lichen-heath that was well-drained.

Mendez *et al*. (1998) demonstrated that α varied markedly over short distances such as a transect across uplands (0.95) and wetlands (1.15). They suggest that α be calibrated to the specific surface type.

Does the Priestley–Taylor equation have any relevance to evaporation from forested areas? Shuttleworth and Calder (1979) examined the relative role of the flux immediately from tree crowns due to interception of rainfall rates; in spruce stands of Wales, it amounts to 27% of the entire evaporative loss.

Indirect measurements at Valday, Russia: a special case study

An Automated System for Data Sampling and Assimilation (ASDSA) was developed by the Institute for Water Automatic at the town of Bishkek (now the capital of Kirghiz Republic) as ordered by the State Hydrological Institute. It was in operation at Valday during three summer seasons (1988–90). The ASDSA sensors were installed above a meadow surface. Computer-controlled, four-level profile measurements of air temperature, air humidity and wind speed were routinely made over one-half hourly increments. The global solar, long-wave and net radiation fluxes were integrated for the same time. Daily evaporation from land and water surfaces were also measured independently using weighing, hydraulic (floating) and pan evaporimeters. To process the data, Penman's equation and the profile method were applied with the restrictions from Equation (6).

By using the ASDSA, short-time variations of atmospheric parameters were observed (Shutov 1998, 2000). Actual individual daily measurements of evaporation reflected the influences of diurnal meteorological variations as they differed from the expected smooth sine curves for daily fluxes. Decreasing net radiation and potential evaporation in the afternoon were due to the variable cumulus clouds growing at that time. Actual evaporation rates decrease gradually, unlike the radiation fluxes. During rain showers, both potential and actual evaporation rates fall to zero, increasing again after the event in spite of diminishing late afternoon solar radiation.

Individual peculiarities in daily evaporation variability are certainly smoothed by in-time averaging of data over several days, thus producing results more similar to the classical sine curve (Figure 2). The standard deviations of $\sigma(E_0)$ and $\sigma(E)$ reflect variable weather

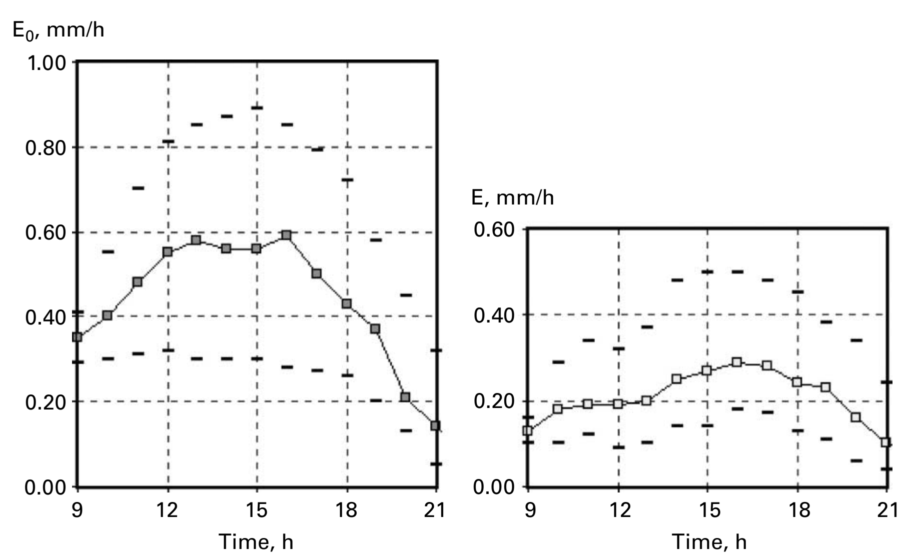

Figure 2 Time-averaged daily courses of potential (E_0, on the left) and actual evaporation (E, on the right) by the ASDSA measurements. Dashes show possible deviations from average

conditions (where E_0 is potential evaporation). Hourly evaporation variability between days increases during the day; and it is greater in the evening hours than in the morning hours, again being affected by summer rainfall from convective clouds.

During the summer season, the daily evaporation rates change from 0.9 to 5.0 mm/d, while potential evaporation rates change from 2.5 to 12.0 mm/d. It was also found from the data analysis that the daily evaporation rates obtained using the ASDSA and using weighed soil lysimeters are poorly correlated.

Calculation methods accepted in Russia: a brief review

The so-called complex method has been developed (Budyko 1971) to calculate evaporation from grassland surface based on the air humidity deficit and soil moisture content at the top of the soil layer (observed or computed). Accepted variants (Balonishnikova *et al.* 2004) are as follows:

$$E = \frac{E_0 W_{av}}{FC} \tag{9}$$

where E_0 is possible potential evaporation (mm), W_{av} is soil moisture content in the top soil layer 1 m deep averaged for a period of time (mm), and FC is the field capacity of soil (mm).

To determine potential evaporation, a graphical dependence between E_0 and the water vapor pressure deficit was applied.

A method developed in Valday has been recommended to determine evaporation from a forested watershed where systematic measurements are absent (Fedorov 1978). Evaporation from a forest for the warm season is calculated according to the following equation:

$$\sum E = A \sum E_0 \tag{10}$$

where A is an empirical factor dependent upon the "radiation humidity index" (Budyko 1971), calculated as $\Sigma Q_{net}/L\Sigma P$. Monthly potential evaporation E_0 and net radiation Q_{net} divided by the latent heat of fusion and total monthly precipitation P are calculated for each month. An empirical dependence of the factor A upon the humidity index is used.

Evaporation-soil moisture relationship

Consider the problem of how to determine the key factor α in the Priestley–Taylor equation; should it be a constant or variable for a specified region, landscape, vegetation, etc? The variable factor approach is suggested here, depending on soil moisture content SM, i.e. $\alpha = f(SM)$ (the function can often be non-linear). This non-linearity was observed much earlier in work by Denmead and Shaw (1964) and is likely to be due to the non-linear variation in soil particle/moisture surface tension relations. In this case, the method becomes more universal, resembling the so-called Heat-and-Water Balance (HWB) method proposed in recent Russian investigations (Shutov 1998, 2003).

The HWB method allows us to model all the basic conditions of evapotranspiration: meteorology, hydrology and plant physiology (respectively represented by the potential evaporation, soil moisture and an empirically fitted factor related to plant species and their growth stage). The HWB scheme is based upon the following equation:

$$E = \beta E_0 F(SM) + kP \tag{11}$$

where β is the factor of plant transpiration activity determined from long-term lysimetric observational data (Shutov 2000) and, probably, correlated with the Leaf Area Index (LAI), E_0 is the potential evaporation (defined as mentioned above by Budyko (1971)), and $F(SM)$ is a function describing the influence of soil moisture given in an empirically based equation. The k factor simulates the effect of rainfall interception as seen in observations at a grassland site in Valday that reached 0.22 by late summer and early fall. All terms in Equation (11) are in mm.

Soil moisture data has been obtained using the neutron probe method (Kapotov and Shutov 1993). It can also be obtained by remote sensing methods such as satellite-based synthetic aperture radar (SAR) (Meade *et al.* 1999). The soil moisture is highly variable and is one of the main causes, along with radiation fluxes, for the spatial variability of evaporation. The corresponding relationships have long been used in many calculation techniques (Hansen and Jensen 1986).

Measurements for a number of seasons at Valday (Shutov 1998) allow one to find the relationship between evaporation and soil moisture (Figure 3). We have examined the ratio of actual evapotranspiration to evaporation from open water (which is assumed to be near to potential evaporation) and the ratio of actual soil moisture SM to the field capacity (FC) of the soil, which limits the capillary flux contributions to evaporation. In spite of scattering, one can see that the relationship appears non-linear, particularly for conditions when $SM/FC < 0.6$, i.e. for droughts. For SM/FC ranging from 0.20 to 0.98, the $F(SM)$ curve can be represented as a power function:

$$F(SM) = W_c^* \left\{ 1 - \left[1 - (W^*)^{1/m} \right]^m \right\}^2 \tag{12}$$

where the exponent m lies between 0.51 and 0.88, depending on the soil textural type. The function corresponds to that describing hydraulic conductivity $K(W)$ of the water flux within soil profile (Shutov and Kaljuzhny 1994). This complicated power equation was derived from an analytical description of soil as a porous medium (van Genuchten 1980). The simple linear HWB schemes developed earlier may be assumed correct only when dealing with high soil moisture, $W/FC > 0.8$. Conversely, when $W < 0.7$ the $F(W)$ curve can only be applied through a non-linear function.

Errors and shortcomings: general outlook

The shortcomings of direct lysimetric observations seem clear: this method underestimates evaporation in most climatic conditions. Systematic errors of the lysimetric method are caused mainly by restriction of free water transfer in the soil profile which is confined within the lysimeter. These errors depend on climate wetness conditions and ranged from −4% when potential evaporation E_0 is equal to precipitation amount, i.e. $E_0 - P = 0$, to −18%

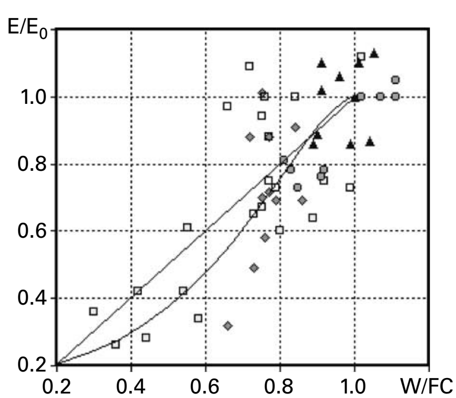

Figure 3 An empirical evaporation–soil moisture relationship obtained at Valday, Russia. The straight line is simply a 1:1 fit and the curved line is a best fit

when $E_0 - P = 500$ mm. So, if we assume $E_0 - P = 125$ mm (that is, the maximum value for the Alaskan North Slope), the inaccuracies are about -8% for lysimetric observations.

The energy balance method usually overestimates evapotranspiration rates; the error is rather large for drought conditions. For instance, if $E_0 - P = 125$ mm, the corresponding relative error is $\Delta E/E = +10\%$ and for the annual wetness deficit $E_0 - P = 375$ mm (this is similar to the Central Canadian Plains) where the error is determined as $\Delta E/E = +22\%$. One of the main sources of error in ET estimates, as well as spatial variability in evaporation rates, is the highly variable soil moisture content. Variations are greatest in early spring and autumn because of an uneven distribution of the snow pack in the first case and the effect of spatial non-homogeneity in the soil water properties in the second case. Attempts to consider these effects were addressed in the relationship above between evaporation and soil moisture content (Shutov 2003). Of particular interest are the scaling effects in soil moisture variability, which can modify those areal evapotranspiration estimates obtained through the use of land-based or remotely sensed data (Georgiadi *et al.* 1990; Serafini 1990).

A direct inter-comparison of all the methods for determining evapotranspiration was carried out based upon the long-term time series data obtained at Valday. We have tested some of the methods by using a model (normal) year, when observations were made in full. Results obtained (Figure 4) show that both calculation schemes (Budyko 1971; Shutov 1998) overestimate evaporation in May. The empirical relationship for potential evaporation described here (Budyko 1971) ignored radiation effects and therefore is insufficiently proven. Lysimetric observations gave evaporation rates less than those acquired by the energy balance–Bowen ratio method due to disrupted water exchange within lysimeters, a limitation discussed above.

Comments on several case studies

Following are comments from case studies presented in the NRB workshop noted above and published in *IAHS Publ.* 290 (Kane and Yang 2004). The case studies have been selected to outline several problems with the observation and calculation techniques, to examine data sets for several climatic conditions and to exemplify several issues related to the annual cycle of evaporation, temporal variability, local (e.g. plant species, elevation) influences, etc.

Hydrological cycle on the North Slope of Alaska, USA (Kane *et al.* 2004)

Evapotranspiration (ET) over the basin has been determined as a residual: $ET = P\text{-}R$ and observations were also carried out for evaporation from water surface (EW) with the use of a

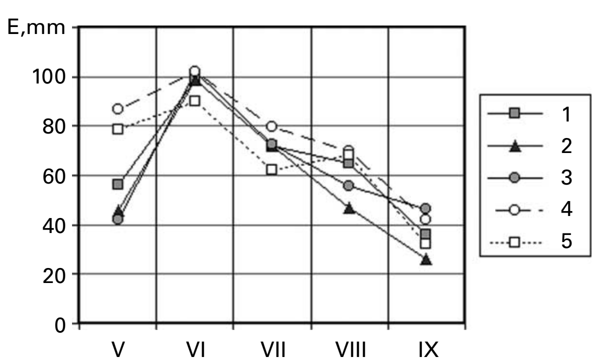

Figure 4 Monthly evaporation rates for a summer season at Valday. 1- water surface evaporation as measured with the use of 20 m^2 area basin, 2 - grassland evapotranspiration measured directly by weighing lysimeters, 3 - grassland evapotranspiration determined by the energy balance method, 4, 5 - evapotranspiration calculated by Budyko (1971) and Shutov (1998) respectively

class A pan evaporimeter. For this watershed, we can see the quite satisfactory time series. The available data enable analysis of trends and irregular fluctuations (Figure 5). It is notable that there appears to be a decreasing evaporation trend (about in 40 mm a decade) from the pan (although not evident in estimates of *ET* from water balance). This is similar to what has been observed at Valday, Russia. What could cause this trend? In our opinion, it might be a recent increase of air humidity, decreasing the vapor pressure gradient. Might it be due to an effect of wind speed or changes in meteorological conditions? There is no clear answer to this question presently.

Water balance dynamics in Interior Alaska (Bolton *et al.* 2004)

Calculation of evapotranspiration was based upon the Priestley–Taylor equation. The method was chosen due to the minimal amount data input required. The factor α was assumed to be 1.13 (Rouse 1976), and $s/(s+\gamma) = 0.406 + 0.011T_a$. The calculated E (in mm) was then related to the maximum daily air temperature (T_m) by the following empirical function:

$$E = -0.407 - 0.001T_m - 0.003T_m^2. \tag{13}$$

The correlation of this relationship is too low ($r = 0.71$) to use Equation (12) confidently. It seems one may confidently accept the equation only for $T_m < 15°C$, where it is more reliable. Beyond the limit (say, at 18°C), the predicted E values scatter from 0.2 to 3.7 mm/d (see Figure 8, Rouse 1976). Determination of a constant α factor remains problematic.

Water balance at the Mittivakkat glacier basin, Greenland (Hasholt and Mernild 2004)

Due to its simplicity, an early formula offered by Makkink (1957) was applied for the reference evaporation. Actual evaporation was calculated using the SnowTran-3D model developed by Liston and Sturm (1998) for a single winter.

Originally, the Makkink formula was derived from Penman's equation and in its simplified version is as follows (Brutsaert 1982):

$$E \cong \frac{0.61Q_{net}s}{L(s+\gamma) - 0.12} \tag{14}$$

where the definitions of R, s and γ are the same as for Equation (8).

For comparison, our recent studies (Shutov 2003) have shown that evaporation from a water surface can be assumed to be a good proxy for the potential evaporation and is proportional to net radiation: $E_w = 0.61R/L$.

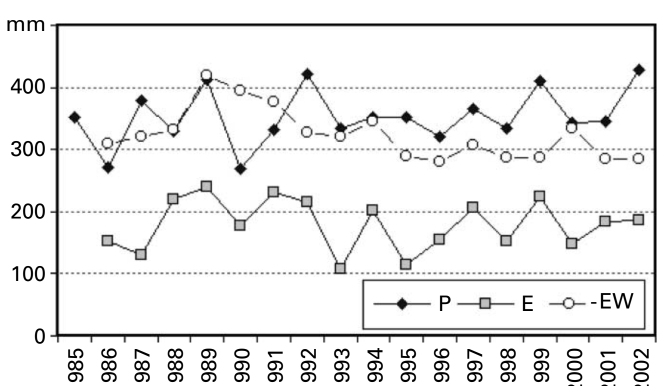

Figure 5 Precipitation (*P*, mm), water surface evaporation (*EW*, mm) and land surface evaporation (*ET*, mm) on the North Slope of Alaska (Kane *et al.* 2004) during the warm season for the period from 1985 to 2002

Generally, the calculated evaporation from the Mittivakkat basin is unexpectedly high. It is questionable whether there are such high evaporation rates during April to August. These rates are as high as in much lower latitudes (e.g. at Valday). Probably, there is a need to re-fit the factors in Makkink's formula for the Greenland data because the constant was obtained from observations in the Netherlands. It may be necessary to use a reduced factor instead of 0.61 when applied to hydrological studies on Greenland.

Very long-term studies at Valday, Russia (Balonishnikova *et al.* 2004)

Two studied watersheds, Log Usadjevsky (grassland) and Log Taezhny (boreal woodland) are located on the Valday Hills in north-western European Russia. This area is characterized by a continental wet climate. Mean annual air temperature is 3.1°C. The warm season is 212 d long. Mean annual precipitation at Valday meteorological station equals 780 mm, of which 70% is rainfall. Snow fall and the snow cover occur annually and are of great importance to the area.

The most valuable aspect of this study is that all the water balance components have been determined independently for more than 30 years, with considerable attention given to the measurement accuracy. Therefore, the data obtained are a basis for the study of seasonal partitioning between several water balance components (Figure 6).

It is interesting to note that summer evaporation from woodlands increases significantly (~25%) due to forest transpiration, thus total *ET* loss exceeds precipitation. Additionally, there is considerable loss in soil moisture due to water uptake by root systems. Inter-annual variability of evaporation rates is quite small (variability factor $CV = 0.08$ for the summer time) in comparison with other high latitude (e.g. Alaska) regions.

Evaporation from boreal wetlands, Northern Russia (Shutov 2004)

Wetland, moss, fen and bog: the question is whether they are really typical of intact landscapes of the sub-Arctic? Evaporation from wetlands in northern Russia was measured using a pan evaporimeter (for open water) and specially constructed weighable lysimeters, which were installed in pairs (at different micro-relief elements). Evapotranspiration rates from wetland landscapes are unexpectedly lower (57–63%) than those from an open water surface. The second issue is that evaporation from wetlands are closely correlated with solar radiation, thus allowing for a convenient approach to indirectly estimate *ET* losses.

Evapotranspiration rates are influenced by decreasing ground water levels (GWL); lower water tables are produced when the top soil layer is highly porous. These organic soils have no fine pores, so the capillary fringe is small. When GWLs become deeper than approximately −50 cm, a significantly dry condition may occur for those dwarf plants native

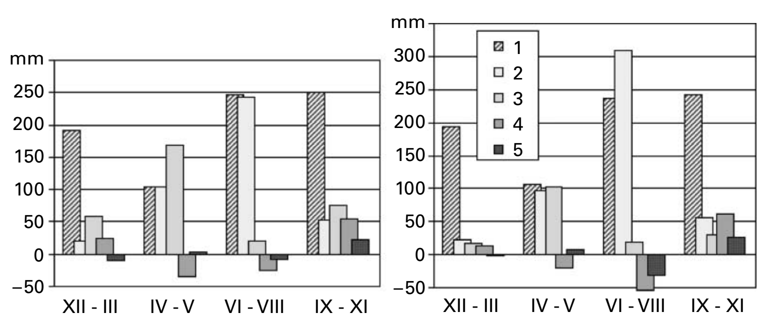

Figure 6 Seasonal water balance components at Valday, Russia, observed for grassland (on the left) and forested watersheds (on right, Balonishnikova *et al.* 2004). 1 - precipitation (*P*), 2 - evapotranspiration (*ET*), 3 - runoff (Q_r), 4 - soil moisture change (ΔSM), 5 - ground water storage change (ΔGW)

to wetlands. This accounts for two geo-botanical features: first, each of the GWL ranges corresponds to several moss species, and second, the specific dwarf shrubs are xerophytes (rough leaves, etc.).

Experimental hydrology studies at Mogot, Eastern Siberia, Russia (Vasilenko 2004)

The site location is properly beyond the sub-arctic zone, but it characterizes the Siberian *taiga* with its permafrost and extreme continental climate. It is an area of low mountains with an elevation range between 500–1200 m a.s.l. Evaporation was determined using three different methods simultaneously: energy balance (standard), with assumption of the Bowen ratio and modified, and direct lysimetric observations.

Annual evaporation values ranged from 258–420 mm, showing substantial inter-annual variability. Negligible winter evaporation (along with relatively low snow pack accumulation) and significant summer values (Figure 7) are the most distinctive features of the continental climatic regime. It is notable that spring runoff is practically absent due to low snow and considerable evaporation by March and April. Summer runoff is generated by persistent rainfalls, partially of monsoon origin.

Investigations made near Tiksi, Russia (Ishii *et al.* 2004a)

Potential evapotranspiration E_0 over the wet moss was calculated by the Penman method in 1997. The Bowen ratio method was used in 1998 and 1999 to estimate actual evapotranspiration over the wet moss E_w. According to Sato *et al.* (2001), $E_w = 0.71\ E_0$. Relationship between evaporation from gravel and wet moss is as follows: $E_d = 0.18E_w$, hence $E_d = 0.128\ E_0$. Partial areas of the watershed under study are 0.64 (wet moss) and 0.36 (gravel land). The water balances for three summer seasons (Table 1) demonstrate significant variability in precipitation (from 90–217 mm) and evaporation (from 33–75 mm). It resembles the highly variable evaporation rates for Alaska's North Slope (see Figure 5). Perhaps this may be a common climatic feature of low Arctic and sub-Arctic regions, where precipitation regimes can often alternate year-by-year between humid and very dry conditions.

Kolyma case study, North-eastern Siberia, Russia (Zhuravin 2004)

The site lies in the mountainous taiga zone of the Northern Hemisphere with continuous permafrost, with extremely cold (−43°C) winters. Quite distinctive is the relatively low evaporation (only 40% of precipitation totals) and this can be explained by the high amount of conductive energy flux (Q_c) into the frozen soil during the summer period (Table 2). The low mountainous experimental watersheds include also the so-called *golets*: these are uplands and ridges covered with coarse gravel and boulders where evaporation is strongly limited.

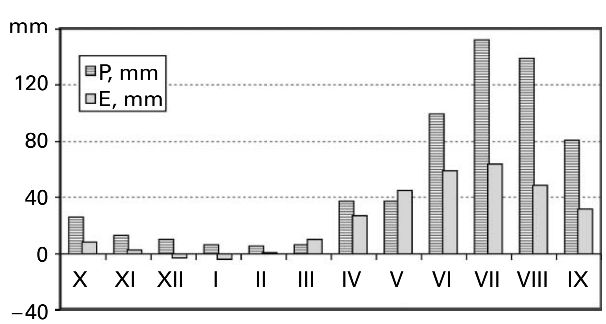

Figure 7 Pattern of monthly precipitation (*P*) and evaporation (*E*) at Mogot, East Siberia (Vasilenko 2004)

Table 1 Water balance components (mm) at Tiksi, North Siberia (Ishii *et al.* 2004a, b)

Year	Period	*P* (mm)	*M* (mm)	*E* (mm)	*R* (mm)
1997	23/VI–31/VIII	217	130	75	338
1998	5/VI–26/VIII	90	129	33	186
1999	5/VI–31/VIII	98	83	54	144

Research basin studies in Yukon, Canada (Janowicz *et al.* 2004)

Another example of semi-arid, continental area is the Canadian Yukon, where observations were made in three particular ecosystems according to three elevation zones. Evapotranspiration was determined by Morton's method, which likely gives overestimated values, since the ground heat flux was not taken into account. Boreal forest, sub-alpine shrubs and alpine tundra cover 22%, 58% and 20%, respectively, of the watershed area. On average (Table 3), the evaporation from boreal forest is 1.34 times that in sub-alpine shrubs and 1.70 times that in alpine zone. Local climate and permafrost conditions are milder here than the Kolyma in Russia and therefore higher evapotranspiration rates.

Water balance of a snowy watershed in Hokkaido, Japan (Ishii *et al.* 2004b)

Penman's method was used for evaluating evapotranspiration after 1992. Previously, the Thornthwaite method (Thornthwaite 1948) was used because there were no net radiation data available. Penman's equation for evapotranspiration (E_p) was used in the following form:

$$LE_p = \frac{s\left(Q_{net} + F(u)\left[e_s^*(T_a) - e\right]\right)}{s + \gamma} \tag{15}$$

where $F(u) = 0.26(0.5 + 0.537u)$, and u = wind speed (m/s) at 1.5 m level.

Heat flux into the soil Q_c was assumed to be negligible. Takeuchi *et al.* (1994) reported that daily evaporation over a grassland measured by a weighing lysimeter was close to the E_p determined by Penman's formula. Therefore E_p was considered as true evaporation. This is, of course, always a disputable assumption in most regions, but it is reasonable in the Japanese Islands with their very humid climate conditions. Comparisons of the evaporation estimates by Penman (E_p) and Thornthwaite (E_T) methods for July through December gave $E_{Pen} = 1.24E_{Thorn}$ ($r = 0.96$), and for January through June, $E_{Pen} = 0.85E_{Thorn}$ ($r = 0.91$).

Canadian Arctic: a case study at the highest latitude (Young and Woo 2004)

In contrast to the Moshiri basin in Japan, the highest latitude studies show the lowest evaporation rates. These rates were determined in an area of polar oasis using the Penman–Monteith method (Woo and Young 1997). They found significant variations caused by elevation gradient and summer conditions; daily evaporation rates ranged from 0.45 mm to 1.23 mm. Averages for the warm season at 76°N (Ellesmere Island) reaches 63–81 mm. For

Table 2 Water balance components (mm) at Kolyma, North Siberia (Zhuravin 2004), including totals for the warm and cold seasons

Component	Cold	V	VI	VII	VIII	IX	Warm	Year
P	135	35	65	96	84	55	336	471
E	9	17	34	36	27	14	128	137
R	4	37	96	72	59	28	292	296

Table 3 Evaporation (mm) from various ecosystems at Wolf Creek, Yukon, Canada (Janowicz *et al.* 2004)

Ecosystem	V	VI	VII	VIII	IX	Total
Alpine tundra	48	55	56	37	14	211
Sub-alpine forest	64	69	66	49	20	268
Boreal *taiga* forest	71	90	91	71	36	359

a patchy wetland on Cornwallis Island at practically the same latitude, the Priestley–Taylor method gave up to 4–6 mm/d, i.e. the values which are quite comparable to those observed in the sub-Arctic; so the total would reach 145 mm for the summer.

Summary of regional evaporation rates

Evaporation rates have never been examined for the entire Arctic and sub-Arctic regions. One of the earlier works (Mendez *et al.* 1998) summarized the data compiled from northern North America. That summary has become a basis to evaluate the evaporation rates expected for Arctic regions. The monthly and seasonal values can be significant, namely up to 330 mm for the warm season from May to August. By having access to the evaporation data that are representative of the pan-Arctic, we can outline the regional rates more widely. This has been done and is presented here in Table 4.

Of course, we cannot consider these data as homogeneous and quite comparable as related to different time intervals. However, some preliminary analysis may be done in the form of the relationships between evaporation, precipitation and the radiative energy. This approach assumes that the annual evapotranspiration E, on the average, can be evaluated through the following theoretical Equation (Brutsaert 1982):

$$\frac{E}{P} = 1 - e^{(Q_{net}/LP)}. \tag{16}$$

Here Q_{net} is net radiation (MJ/m^2 per year), LP is the energy needed to evaporate the precipitation amount P and the ratio Q_{net}/LP is the index of humidity (Budyko 1971).

In a correlation plot of E/P vs. Q_{net}/LP (Figure 8), we relate the data for several northern watersheds. First we should note that the empirical curve (A) slightly diverges from the theoretical one (B). We approximate it with the following formula with correlation rate $r = 0.85$:

$$E/P = 0.377 \ln(Q_{net}/LP) + 0.595. \tag{17}$$

The condition where $Q_{net} = LP$ would signify that the net radiant energy and its latent heat flux that corresponds to evaporative loss of precipitation are equal. In such a comparison, humidity must be assumed to be at the optimum, yielding evaporation that is roughly 60% of precipitation. The remaining water is a source potentially available for runoff. When $Q_{net}/LP > 1$, that signifies insufficient moisture and the ratio of E/P slowly increases to its upper limit of $E/P = 1$ when Q_{net}/LP approaches infinity. This of course is not realistic. When $Q_{net}/LP < 1$, the ratio E/P can approach ~ 0.0 for high values of relative humidity (low vapor pressure gradients).

It is interesting, but perhaps expected, that the observed E/P values for the northern countries are highly diverse, ranging from 0.1 to 0.9 and even slightly more. Obviously, there are highly variable atmospheric moisture and thermal conditions: from extremely humid Baffin Island and Hokkaido Island to the extremely severe cold of Eastern Siberia and relatively dry central Canadian plains.

Table 4 Summary (annual precipitation and evapotranspiration, including sublimation)

No.	Study site	Country	Lat, °N	Landscape zone	*P* (mm)	*E* (mm)
	Europe					
1	Valday	Russia	58	Boreal forest	792/794*	442/497*
2	Brusovitsa	Russia	64	Boreal wetland	685	298
3	Askanjoki	Finland	66	Boreal forest	645	222
4	Iittovuoma	Finland	69	Tundra	574	230
5	Bayelva	Norway (Swalbard)	79	Arctic tundra/glaciers (55% of entire area)	385/965**	46
	Asia					
6	Moshiri	Japan	44	Boreal rainforest	1670	370
7	Mogot	Russia	55	Mountainous taiga	612	290
8	Kolyma	Russia	63	Mount. taiga/tundra	471	178
9	Tiksi	Russia	71	Arctic tundra	317	131
	North America					
10	Dead Creek	Canada (Manitoba)	50	Boreal forest	521	327
11	Wolf Creek	Canada (Yukon)	61	Boreal forest (sub-alpine woodland)	288/387*	268/359*
12	Ft. Simpson	Canada (NWT)	63	Boreal wetlands	421	287
13	Caribou-Poker Creeks	USA (Alaska)	64	Boreal forest	412	202
14	Imnavait Creek	USA (Alaska)	70	Arctic tundra	353	179
15	Baffin Island	Canada	69	Arctic tundra	616	154
16	Devon Isl.	Canada	76	Polar desert	185	117
17	Ellesmere Isl.	Canada	80	Polar oasis	141	83

Notice: *, forested area; **, glacier area

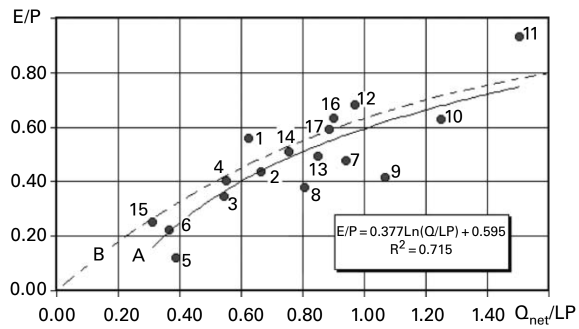

Figure 8 General relationship between the radiation humidity index (Q_{net}/LP) and the ratio of evaporation to precipitation (E/P) for northern research basins. Empirical approximation A (see the formula within inset): theoretical equation B (Budyko, 1971): $E/P = 1 - \exp(-Q_{net}/LP)$. Study sites: 1 - Valday (Russia), 2 - Brusovitsa (Russia), 3 - Askanjoki (Finland), 4 - Iittovuoma (Finland), 5 - Bayelva (Svalbard, Norway), 6 - Moshiri (Japan), 7 - Mogot (Siberia, Russia), 8 - Kolyma (Siberia, Russia), 9- Tiksi (Russia), 10 - Dead Creek (Manitoba, Canada), 11 - Wolf Creek (Yukon, Canada), 12 - Ft. Simpson (NWT, Canada), 13 - Caribou Poker Creeks (Alaska, USA), 14 - Imnavait Creek (Alaska, USA), 15 - Baffin Island (Canada), 16 - Devon Island (Canada), 17 - Ellesmere Island (Canada)

The low values of the E/P ratio on Svalbard and on Hokkaido Island are caused by different effects. In the first case, there may not be a significant evaporation from the non-glacierized part of the catchment (coarse soils and limited vegetation) and on the glacier ice/snow must first melt before evaporating; in the second case, the actual evaporation rate ($E = 370$ mm) is relatively small, but there snowfall is very high and most runoff occurs during snowmelt.

We should also note possible local effects such as chinook, the dry local wind which might cause exceptionally high E/P rate at the Wolf Creek basin in South-western Yukon (just beyond northern Rocky Mountains on the lee side). Most of the conditions examined are grouped roughly within the interval $Q_{net}/LP = 0.8–1.0$ that correspondingly results in $E/P = 0.4–0.65$.

Problems for the future

Evaporation is variable, but predictable in both space and time throughout the pan-Arctic. As hydrological research and field observations continue to advance, we can be more confident in understanding the climatic, biological and geophysical controls on a basin water balance. Evaporation varies predictably, decreasing from south to north (decrease in available energy) and near coastal areas (low vapor pressure gradient). Evaporation is greatest in lower elevations (warmer temperatures). It is greater in the boreal forests than in the tundra regions. It is most strongly controlled by soil moisture levels and incoming solar radiation. Secondary controlling factors include vegetation type, humidity and wind speed.

Only a few data sets exist to enable inter-comparison of various methods of calculating evapotranspiration (Halliwell and Rouse 1989; Kane *et al.* 1990). Such independent data sets, collected in different regions under different conditions, would improve our capability of modeling climate and hydrological processes throughout the Arctic and sub-Arctic. Improvements in the determination of potential evaporation are needed to permit simplified estimations of *ET* to be conducted.

Estimates of evaporation/transpiration over larger areas (compared to point measurements of lysimeters and pans) through new technology would result in better spatial estimates at the catchment scale. Accurate and timely characterizations of soil moisture via remote sensing are also needed to permit better spatial analyses of *ET* variations. Such remote sensing analyses should include dynamic vegetation modeling to characterize spatial and seasonal changes in transpiration factors (such as canopy resistance).

Improved spatial assessments of soil properties (including permafrost extent, texture and organic layer characteristics) are needed to incorporate geophysical controls on evaporation.

This synthesis was not comprehensive in all regions nor did it include all of the tremendous research available, but it does constitute an initial assessment of a circumpolar comparison of evapotranspiration.

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