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Comparative ecosystem-atmosphere exchange of energy and mass in a European Russian and a central Siberian bog I. Interseasonal and interannual variability of energy and latent heat fluxes during the snowfree period

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ABSTRACT

Energy and latent heat fluxes λE were measured over ombrotrophic bogs in European Russia (Fyodorovskoye) and in central Siberia (Zotino) using the eddy covariance technique, as part of the EuroSiberian Carbonflux Project. The study covered most of the snowfree periods in 1998, 1999 and 2000; in addition some data were also collected under snow in early spring and late autumn 1999 and 2000. The snowfree period in Europian Russia exceeds the snowfree period in central Siberia by nearly 10 weeks. Marked seasonal and interannual differences in temperatures and precipitation, and hence energy partitioning, were observed at both sites. At both bogs latent heat fluxes (λE) exceeded sensible heat fluxes (H) during most of the snowfree period: maximum λE were between 10 and 12 MJ m^{-2} d^{-1} while maximum H were between 3 and 5 MJ m^{-2} d^{-1}. There was a tendency towards higher Bowen ratios at Fyodorovskoye. Net radiation was the most influential variable that regulated daily evaporation rates, with no obvious effects due to surface dryness during years with exceptionally dry summers. Total snowfree evaporation at Fyodorovskoye (320 mm) exceeded totals at Zotino (280 mm) by 15%. At the former site, evaporation was equal to or less than precipitation, contrasting the Zotino observations, where summer evaporation was distinctly higher than precipitation. During the entire observation period evaporation rates were less than 50% of their potential rate. These data suggest a strong 'mulching' effect of a rapidly drying peat surface on total evaporation, despite the substantial area of free water surfaces during parts of the year. This effect of surface dryness was also observed as close atmospheric coupling.

1. Introduction

In Russia peatlands are a major element of natural landscapes, covering a total area of 1.54×10^6 km^2 (Kobak et al., 1998). In the present taiga zone, the process of peatland formation after the last glaciation began as early as 8000-8500 years ago, but the development of bogs and peatland formation in woods accelerated ca. 7500-5000 years ago. In western Siberia during the last 8000 years bogs expanded on average over an area of nearly 10^4 ha annually (Piyavchenko, 1980); however, due to drainage and possible effects of climate warming the current situation is less clear.

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Most Russian peatlands are found imbedded in naturally forested regions. There, basically, optimum climatic conditions for peatland formation processes combine with the appropriate geomorphological factors. In the taiga, raised string, *Sphagnum* and blanket bogs make up nearly 44% of the total peatland area. Bogs (and other peatlands) when occupying large areas determine regional values of evaporation and contribute to the regulation of atmospheric air humidity and temperature. They render a significant influence on energy and water balances of entire territories via the circulation of water. In fact, it has been argued that the absorption of net radiation in large wetlands may considerably soften the continental climate of Russia. The elevated relative humidity of surface air that is observed over peatlands may not only be beneficial to the vegetation growing within the peatland itself, but may also be measurable in adjacent ecosystems; via stomatal control of transpiration the higher air humidity appreciably reduces the negative influence of the low precipitation on ecosystem water balances. Peatlands also mitigate peak stream flow and remove suspended sediment; they accumulate, redistribute and recycle significant volumes of fresh water. Thus the large bog area in parts of Russia significantly contributes to the ecological balance in the regional biosphere.

Despite the variety of ecosystem services provided by bog systems, there is only scarce amount of information available on the regulation and the seasonal and interannual variability of energy fluxes. The primary objective in our study was to investigate the seasonality and interannual variation of energy and latent heat fluxes in two oligotrophic bogs during the snowfree period. In a second publication the regulation of CO_2 fluxes in the two sites will be analysed. The sites lie within the East European pine bog Province and the East Siberian bog Province, respectively (Botch and Masing, 1983), and are representative of the one of the largest wetland complexes on earth.

2. Methods

2.1. Site description

Based on the vegetation cover (see below), which corresponds to the phytosociological unit *Oxycocco-Sphagneta*, the two sites compared in this study are both typical ombrotrophic peat bogs of the boreal zone (Wheeler and Proctor, 2000). Their microtopography is characterised by partially inundated *Sphagnum*-dominated hollows and better drained ridges dominated by higher plants. There is no contribution of groundwater to vegetation water supply. However, at Zotino the general topography of the area, with undulating forested sand hills surrounding the bog, suggested contribution of surface runoff after large rainfall events and during spring snowmelt. At Fyodorovskoye the surface is truly raised above the surrounding terrain.

2.1.1. Fyodorovskoye. This site is ca. 300 km WNW of Moscow within the Central Forest Reserve (56°27′N, 32°55′E). The measurement tower was located in the middle of a 4.2 km^2 bog surrounded by a mixed spruce-fir forest. The climate of the area is moderately continental, with an annual mean temperature (1970-1998) of 3.9 °C, an annual precipitation of 711 mm, and a May to September precipitation of 397 mm. Snowmelt generally occurs in late March or early April, while first snowfall usually takes place in early to mid-November. The bog's vegetation is dominated by *Sphagnum* ssp. growing in inundated hollows, while ridges that project above the free water surface are dominated by ericaceous shrubs *(Vaccinium microcarpum, V. uliginosum)* and other herbaceous plants like *Eriophorum vaginatum, Rubus chamaemorus, Carex pauciflora.*

2.1.2. Zotino. This site is located ca. 30 km inland from the village of Zotino at the western bank of the Yenisey river (60°45′N, 89°23′E). The measured bog occupies an area of ca. 5 km^2 and is surrounded by mono-specific *Pinus sylvestris* forest, growing on sandy soils. Annual median temperature measured in the nearby town of Bor (61°6′N, 92°1′E) was -1.5 °C between 1960 and 1989; annual precipitation was 593 mm, 45% of which (267 mm) fell during the growing season (Arneth et al., 2001). During the three years of measurements, snow melted rapidly in early May, while first snow fell in late September. Approximately 60% of the bog surface is characterised as hollows. These are made up by *Sphagnum* ssp. lawns and inundated to a seasonally varying degree. Some vascular plant species grow in the hollows, mainly *Scheuchzeria palustris, Carex limosa* and *Andromeda polifolia.* 40% of the bog's surface area is 30-50 cm tall ridges, formed by *Sphagnum* peat. Atop of the ridges the majority of the vascular plants grow, mainly *Chamaedaphne calyculata, Andromeda polifolia, Eriophorum vaginatum* and *Ledum palustre* (N. Savushkina, personal communication). Growing along the ridges few small (<1.5 m) *Pinus sytvestris* trees are found.

2.2. *Eddy covariance measurements*

Measurements of CO_2 and H_2O fluxes, and associated environmental parameters, were conducted in Fyodorovskoye between 13 June and 12 October 1998 and 26 March and 20 November 1999. Measurements were also conducted between 5 April and 6 November 2000, but due to problems with data collection at Fyodorovskoye in 2000 these data are not presented here. For Zotino, reflecting the generally shorter growing season at this more eastern and northern site, measurements were made between 11 June 1998 and 5 October 1998, 10 April 1999 and 15 November 1999, and 1 April and 18 October 2000. Generally speaking both systems were running continuously during these periods, and short gaps in the record of meteorological parameters, due to instrument failure or maintenance, were filled using data from the flux measurement tower in the nearby forests (Milyukova et al., 2002; Tchebakova et al., 2002). Gaps in the eddy flux records were filled after establishing relationships between the measured fluxes and the forcing climatic parameters (see Results).

The eddy covariance systems employed at both sites were, in essence, similar to those used during the Euroflux project (Aubinet et al., 2000). Briefly, a three-axis sonic anemometer with an omnidirectional head (Solent R3, Gill Instruments, Lymington, UK) was installed in 6 m height (Fyodorovskoye) and in 5.6 m height (Zotino) atop aluminium towers. The instrument provides high frequency measurements (20 Hz) of the three components of wind speed and of the air temperature. For measurements of CO_2 and water vapour concentrations, air was drawn from an inlet atop of the tower through a 1/8″ inner diameter BEV-A-LINE tubing to a closed-path infrared gas analyser (IRGA; LI-COR 6262, Lincoln, NE USA) located close to the bottom of the tower in an insulated wooden shelter. The suction pump was placed before the analyser gas-inlet such that the air was pushed through the instrument with a flow rate of approximately 7 L min^{-1}. A pressure transducer (PTP101B, Vaisala, Helsinki, Finland) in the reference cell provided the necessary information to correct the measurements for variations associated with pressure fluctuations created by the pump. The analyser was run in absolute mode with CO_2 and water-free air circulating in the reference cell, using a combination of magnesium perchlorate and soda-lime. Calibration of the instrument was checked regularly once a week using air of known CO_2 (pressurised bottle) and H_2O (dew-point generator, LI 610, LI-COR, Lincoln, NE USA) concentrations. Power was supplied by an array of solar panels located ca. 500 m (Zotino) and ca. 50 m (Fyodorovskoye) away from the flux measurement towers. During prolonged periods of heavy cloud and very early and late during the vegetation period additional power was supplied via fuel generators, located >200 m from the eddy flux systems.

The voltage output from the solent and the gas analyser were digitally synchronised using the the sensor input unit provided with the sonic anemometer. Calculation of sensible heat (H) and latent heat (λE) fluxes were performed online (Kolle, personal communication). The calculation included coordinate rotation to remove possible errors due to sensor tilt relative to the bog surface (Aubinet et al., 2000). Accounting for the time lag between measurements of w' (via the sonic) and x' (via the gas analyser) the signals were digitally synchronised by maximising the covariance. Fluxes were calculated as the half-hourly averages of $w'x'$. Half-hourly fluxes as well as raw data were stored on hard disks and were regularly transferred onto CD-ROM.

Associated with the use of closed-path analysers will be flux losses attributable to the incomplete spectral response of the analyser, to the dampening of the signal along air flow through the tube and to the separation between the tube inlet and the sonic anemometer head. These can be accounted for by comparing normalised cospectra of sensible heat flux to cospectra of the measured latent heat or carbon dioxide flux. The H-cospectra represent the entire turbulent flux-density; there are no losses because of the high-frequency response of the solent and the lack of sensor separation and air suction in a tube. In particular, they are defined by a straight line of slope -4/3 in the inertial, high-frequency subrange. The flux losses associated with measurements of CO_2 and H_2O concentration in the gas analyser can be corrected for by using an inductivity factor which is derived by aligning the λE and CO_2 cospectra with that of H (Eugster and Senn, 1995). At Zotino, the inductivity values were 0.13 for CO_2 during the entire period of measurements and 0.25 and 0.27 for H_2O in 1998-99 and 2000, respectively. At Fyodorovskoye, the inductivities were 0.13 for CO_2 and 0.30 for H_2O in 1998, and 0.12 for CO_2 and 0.23 for H_2O in 1999-2000. These values were derived, and monitored on a monthly basis, by regularly calculating average co-spectra over 5-6 h periods. The corrections typically increased λE by 12-15% and CO_2 by 5-10%. Corrections due to air pressure differences in the gas analyser's sampling cell and in the atmosphere were calculated online by using the built-in pressure transducer. Water-vapour dilution was corrected for by the internal software of the IRGA.

2.3. *Associated environmental factors*

Radiative flux measurements included total downward and upward radiation using a pyradiometer (LXG055), shortwave downward and upward radiation using a pyrranometer (CM14, Kipp and Zonen, Delft, Holland), shortwave downward diffuse radiation using a pyrranometer with a regularly adjusted shadow-ring (CM11, Kipp and Zonen) and photon flux density (LI-190 SA, LI-COR, Lincoln, NE USA). Additional measurements included air temperature (HMP35D, Vaisala, Helsinki, Finland), air humidity (HMP35D, Vaisala, Helsinki, Finland) and wind velocity (A100R, Vector Instruments). These sensors were installed below the sonic anemometer on a boom with exception to diffuse radiation measurements with the shadow-ring installed at ground level 20 m south of the tower. For both sites at ground level a rain gauge was also installed (#52203, Young Instruments, Traverse City, MI, USA) located in close proximity to the eddy flux tower. Soil heat fluxes were measured at five locations at each site with heat flux plates (Rimco HP3/CN3) installed at depth of 0.05 m. For measurement of soil temperatures, platinum resistance thermometers (Geratherm, Geschwenden, Germany) were installed in two locations close to the towers at depths 0.05 m, 0.15 cm, 0.50 m and 1.0 m, respectively. At both sites the environmental data were collected every 10 s and stored as 10 min averages on data loggers (Campbell CR21X and D13000, Delta-T, Burwell, UK). For comparison with half-hourly eddy flux data, 30 min averages of the environmental data were subsequently calculated.

2.4. *Energy balance*

The degree of energy balance closure is one indication of the performance of eddy flux systems. However, for markedly heterogeneous surfaces such as the two bogs compared in this study, the accurate measurement of components contributing to the energy balance with the same effective footprint is hardly possible. Particularly difficult are representative measurements of soil heat flux density, G, although this flux can be expected to constitute a large component of the surface energy balance due to the lack of above-ground vegetation and because of the large storage of heat in the water: Representative measurements in ridge-hollow complexes with varying degrees of vegetation cover, pure peat and free water surfaces require an unreasonably large amount of sensors. Additionally, the thermal conductivity of the soil heat flux plates generally is much less than the thermal conductivity of the peat-water mix. Moreover, air gaps may develop around the soil heat flux plates as the peat surface dries during summer (Lafleur and Rouse, 1988). A third problem is introduced by the free water surfaces in the hollows, which are of varying area throughout the growing season: a fact that makes not only measurements of G but also representative measurements of net radiation (R_n) difficult. As a consequence it is nearly impossible to measure available energy, $R_a(=R_n - G)$, in peatlands correctly.

To elucidate underlying factors the energy balance was more closely investigated for Zotino site: there, after measured G were corrected for the rate of heat stored in the top 5 cm soil layer above the plates (Campbell, 1985; Halliwell and Rouse, 1987; Lafleur and Rouse, 1988; Moore et al., 1994), half-hourly R_a and $H + \lambda E$ differed by between 15 and 35% (grouping the data into monthly bins). The estimation of the heat-storage term included calculation of $\Delta T/\Delta t$, the rate of change of temperature over the top 5 cm. This in turn was averaged from surface temperature, 1, 2 and 5 cm, reflecting the exponential T-decline with depth. Temperatures at 1 and 2 cm were calculated from surface (pyrradiometer) temperature using a damping depth of 0.05 m (Monteith and Unsworth, 1990; Campbell and Norman, 1998). Assuming a ratio of 80:20 for water and peat on a per volume basis, the thermal conductivity of the soil was an estimated 0.51 Wm^{-1} K^{-1} (Monteith and Unsworth, 1990), nearly 30% larger than the thermal conductivity of the heat flux plates given by the manufacturer.

3. Results

3.1. *Climate*

Figures 1 and 2 summarise measured 1998—2000 daily average air temperatures, surface albedo and

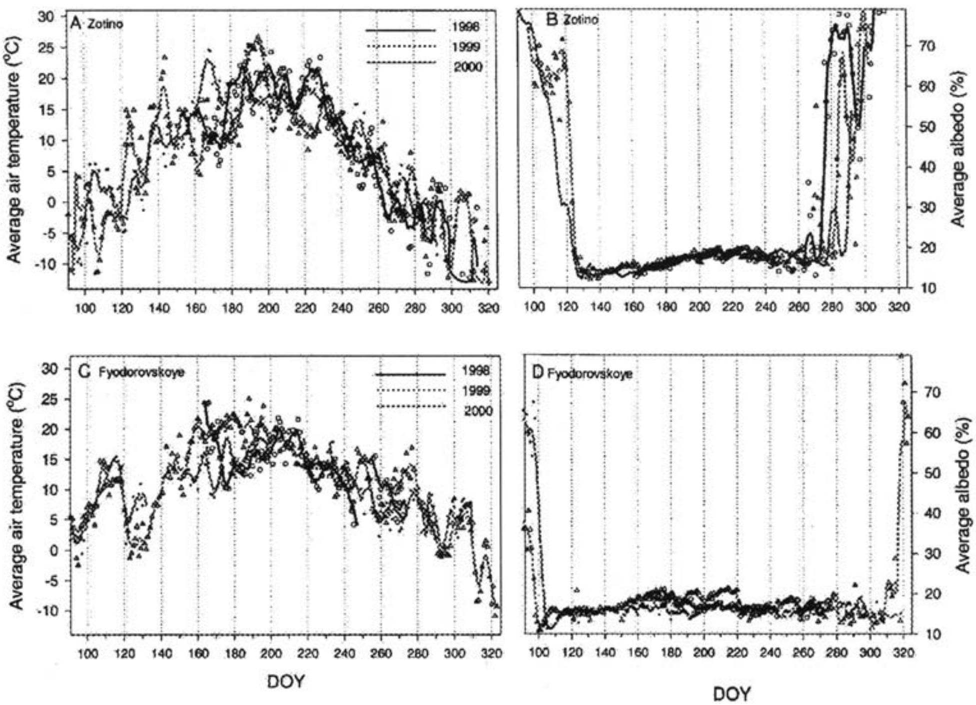

Fig. 1. Time courses of climatic data measured at the central Siberian (Zotino) and European Russian (Fyodorovskoye) bog compared in this study. DOY = day of the year. A and C show daily air temperatures in 1998 (circles), 1999 (triangles) and 2000 (dots) between early April and late November. The lines represent 5-day running means through the data. B and D compare daily average surface albedo at the two sites. Symbols and lines are as in A and C.

precipitation for both sites (at Fyodorovskoye only 1998 and 1999).The lines in Fig. 1 represent 5-day running means through the data.

As expected from its more continental location, spring temperatures at Zotino increased much later and autumn temperatures decreased earlier than at Fyodorovskoye (Figs. 1A and C; Table 1). As indicated by rapidly declining surface albedo values, snow melted in late March/early April over a period of approximately 15 days at Fyodorovskoye, whereas snowmelt at the central Siberian site did not begin until early May. Snow melt happened generally more rapidly at Zotino, especially so in 1999 (Figs. IB and D). There was considerable year-to-year variability in the date of the first snowfall at Zotino, which was observed in late September in 1998 and 1999 but only in late October in 2000 (Fig. 1). For Fyodorovskoye the site was without snowcover until mid-November in all three years. The snowfree period at Fyodorovskoye thus typically exceeded the snowfree period at Zotino by nearly 10 weeks. At both sites, minimum albedos were measured directly after snowmelt. At Zotino, albedo then increased at a steady rate to maxima around early August, declining thereafter. At Fyodorovskoye, albedo values were much more variable throughout the remainder of the snowfree period and did not show a seasonal trend.

Marked seasonal and interannual differences in temperatures or precipitation were typical for both sites. For example, in early spring 2000, temperatures at Zotino were nearly 10 °C cooler than in 1999 [Fig. 1A,day-of-year(doy) 120-130, 140-150]. Only a few weeks later, in June, the pattern was reversed and air temperatures were 10 °C above the 1998 and 1999 values (doy 160-180). For Fyodorovskoye, early summer in 1999 was exceptionally warm, air temperatures being above the 1998 and 2000

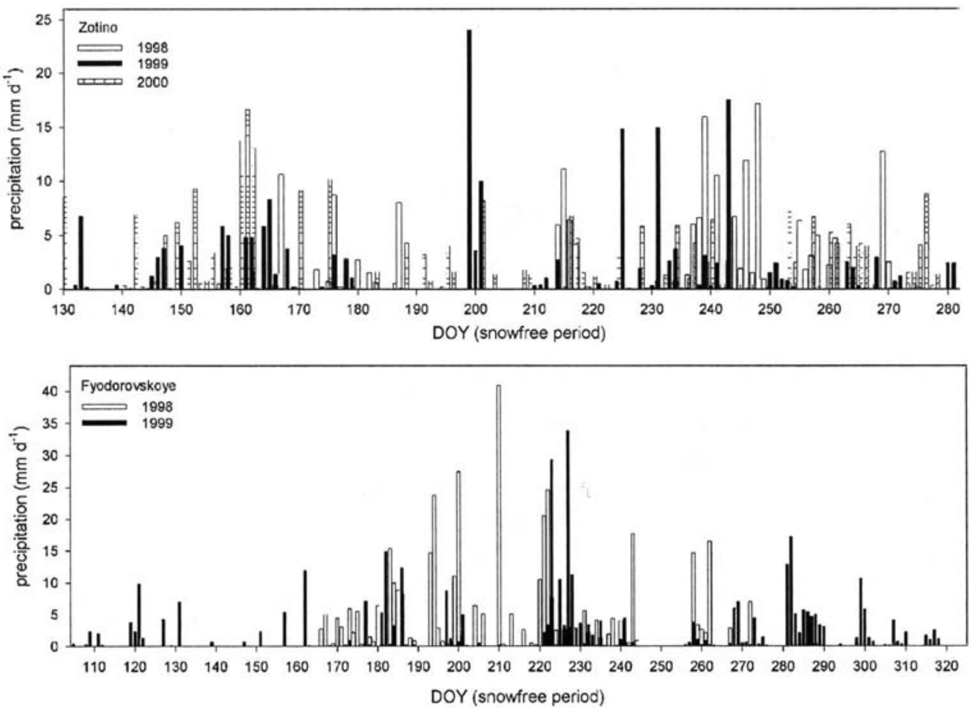

Fig. 2. Daily precipitation during the snowfree period at Zotino (A) and Fyodorovskoye (B). Measurements commenced by mid-June 1998. At Fyodorovskoye data were incomplete during the measurements in 2000, and only data from 1998 and 1999 are shown.

values continuously for nearly one month (doy 150-190).

Rainfall at Zotino during the summer months of 1998 and 1999 was restricted to more or less isolated events separated by periods up to 30 days without any precipitation (Fig. 2A). The driest month of the measurements was July 1998, with only 10 mm of precipitation (Table 1). In 2000, precipitation was distributed more evenly in Zotino, but the single events were of smaller magnitude. In both summers 1999 and 2000 precipitation was below the long-term average measured at Bor. The Zotino and Bor numbers could be compared directly because on average, monthly precipitation measured at the Zotino bog was within 20% of precipitation measured at Bor (Arneth et al., 2002). In Fyodorovskoye, rainfree periods were typically of a shorter duration: generally less than 10 days. Monthly precipitation at Fyodorovskoye typically exceeded that at Zotino by a factor of 2 or more. However, the summer 1999 was exceptionally dry at Fyodorovskoye, and precipitation in June and July was of a similar small magnitude to that observed at Zotino.

3.2. Ecosystem energy fluxes

Figure 3 compares daily integrated R_n, λE and H values for the 1999 measurement period in Fyodorovskoye and Zotino. To aid interpretation, also shown are 5-day running means through the data. In Figures 4 and 5, 5-day running means for all three data periods at the two sites are shown.

At both sites, most of the energy was distributed towards λE for most of the season. Maximum λE values during the summer months were between 10 and 12 MJ m^{-2} d^{-1}, while maximum H were between 3 and 5 MJ m^{-2} d^{-1}. The energy partitioning differed somewhat between the sites such that during summer sensible heat fluxes at Fyodorovskoye generally exceeded H at Zotino (Fig. 3B), while the pattern for λE was reversed (Fig. 3C). At Zotino, ecosystem λE increased rapidly

Table 1. *Monthly average air temperatures (T) and total net radiation (R_n), precipitation (P) and evaporation (E) during the months June-September at the two bogs*[a]

	T, °C	R_n, MJ	P, mm	E, mm
Zotino				
1998-6[b]	12.3	216.1	30	49.5
1998-7	19.1	402.7	10.1	102.4
1998-8	16.6	232.2	73	60.2
1998-9[c]	5.3	49.9	66.1	11.2
1999-6	11.1	305.8	47.3	70.2
1999-7	20.1	377.7	39.2	92.4
1999-8	12.6	200.1	66.2[d]	46.0
1999-.9[c]	5.2	85.3	15.5	22.8
2000 - 6	16.4	371.4	73.2	78.4
2000 - 7	16.4	358.7	29.6	85.9
2000 - 8	15.3	221	46.2	44.9
2000 - 9	7.0	89.7	49.4	21.3
Fyodorovskoye				
1998 - 6[b]	15.8	170.9	39.9	28.6
1998 - 7	16.3	278.9	181.1	58.2
1998-8	13.4	193.3	123.4	44.2
1998-9	13.5	44.0	50.7	-
1999-6	19.2	383.5	30.9	59.4
1999-7	19.1	352.3	47.6	78.3
1999-8	13.9	220.1	116.2	43.3
1999-9	10.1	144.1	25.8	23.4

[a]In 2000, complete data was only available for the Zotino site.
[b]Measurements started on 11 June in Zotino and on 13 June in Fyodorovskoye.
[c]Data until first snowfall; 23 September in 1998 and 24 September in 1999.
[d]Data from a nearby forest flux site (Tchebakova et al., 2002) because of malfunction of the raingauge at the bog.

after snowmelt from near zero to maximum values between 10 and 12 MJ m^{-2} d^{-1} (Figs. 3 and 4). Following the decline in R_n and in temperature (Fig. 1A), and the onset of snowfall (Fig. 1B) in autumn, all of which occurred over a relatively short time frame, λE declined rapidly, being close to zero by late September (doy > 270). Sensible heat fluxes were negative before snowmelt and increased rapidly to maxima as early as late May-June (doy > 150, Figs. 3 and 4). At Fyodorovskoye, reflecting the much earlier snowmelt, the increase of H and λE began in early April (doy > 90, Figs. 3 and 5), while in autumn H and λE declined at a more or less similar rate to that observed at Zotino.

Superimposed on the seasonal trends in the data was a significant scatter in the day-to-day energy exchange: daily sums of λE and H varied by a factor of >2 within only few days and at both sites periods of more than one week could be distinguished with pronounced interannual differences of energy fluxes (Figs. 4 and 5). For example, at Zotino between days 150-160 5-day running means of H and λE in 1999 exceeded the values measured in 2000 by a factor of up to 4. Likewise, in 1998 between days 220-230 λE, but not H, were significantly higher than in the other two years. At Fyodorovskoye in 1998, H were less than in the following two years practically at all times and did not exceed 4 MJ m^{-2} d^{-1} (Fig. 5B).

As reflected in declining Bowen ratios $\beta(H/\lambda E)$ from values greater than 0.7 to values less than 0.3 the partitioning of energy fluxes at Zotino shifted towards a clear predominance of latent heat between early June (doy > 150) to mid-summer (doy > 185). After mid-summer, however, β gradually increased, attaining values typically > 1 around the end of the measurement period in autumn (Fig. 4D). While in early spring and later in autumn on average most of the energy was partitioned into H a clear picture did not emerge. Particularly during autumn when λE were low periods of $\beta \geq 1$ were followed by periods when $\beta < 0.5$, and there was no comparable pattern between

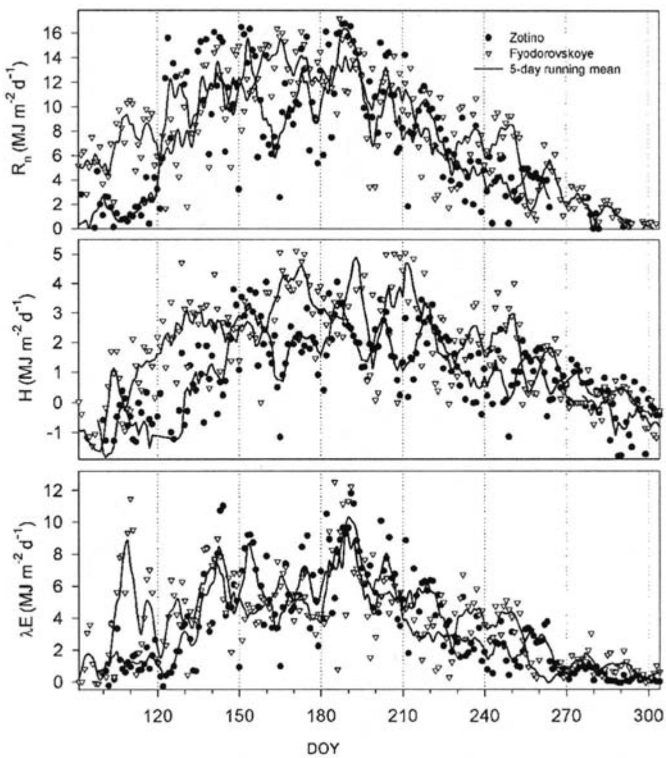

Fig. 3. Daily net radiation (R_n), sensible (H) and latent (λE) heat fluxes and five-day running means (lines) for 1999. Fyodorovskoye data are denoted as open triangles, Zotino data are denoted as closed circles.

years. In both springs 1999 and 2000, β first declined rapidly after snowmelt, then decreased again to values >0.7 before the more general decline over the growing season began (Arneth et al., 2002). This pattern in both years was accompanied by rapidly increasing, relatively warm temperatures (Fig. 1A). A somewhat similar trend of rapidly declining β after snowmelt, followed by increasing values, was also observed in spring 1999 at Fyodorovskoye. However, at this site no seasonal trend in β was observed afterwards. Bowen ratios generally exceeded values in Zotino, and during August 1999 H was frequently equal to or larger than λE, indicated by $\beta > 1$ (Fig. 5D). During most of 1999 β thus exceeded the preceding summer's values, but as for Zotino, β in autumn became very variable.

From Figs. 4 and 5 it is clear that at both sites, fluctuations in λE followed to a large extent the fluctuations in net radiation during the snowfree period (Figs. 6A and B). At Zotino, between 84 and 93% of the variation could be explained by a linear relationship between R_n and λE, which showed a negligible offset. On a daily basis, more than 50% of net radiation was diverted into evaporating water, but the slopes differed somewhat between years. In contrast, at Fyodorovskoye <50% of net radiation was distributed towards evaporation, and the slopes of the regressions were nearly identical but the linear relationship explained less of the variation, particularly so in the dry year 1999 between days 109-111 (triangles) and during doy 185-191 (circles). In both cases λE at a given R_n was exceptionally high. Before snowmelt, as well as on days when average temperatures were less than 5 °C (often associated with night frosts), λE were independent of R_n.

Using the regressions between λE and R_n shown in Fig. 6 to interpolate missing values due to instrument failure, total bog evaporation expressed in millimetres water equivalent (E) between mid-June and

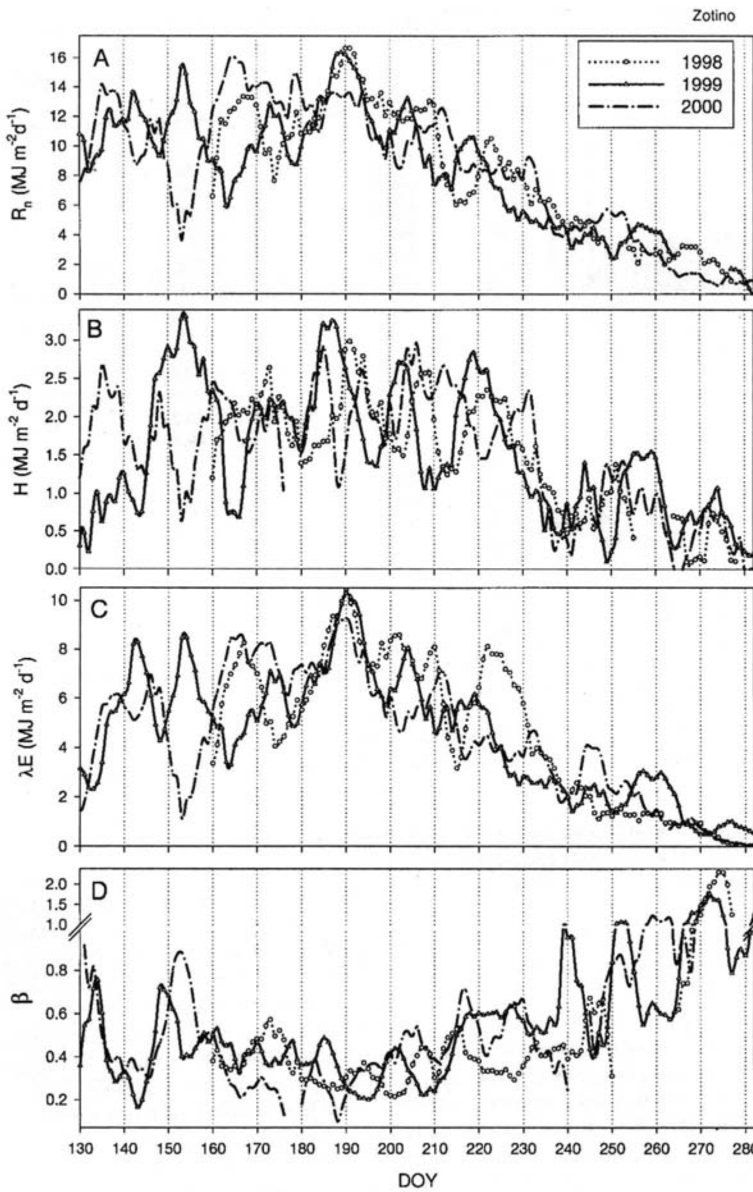

Fig. 4. Five-day running means of daily net radiation, sensible and latent heat fluxes and Bowen ratio at the Zotino site.

30 September at Zotino was 226 mm in 1998, and between 1 June and 30 September 250 mm in 1999 and in 231 mm in 2000 (Table 1). For the entire snowfree period (including evaporation in May), E was remarkably similar, 280 mm in 1999 and 275 mm in 2000. In 1998 and 2000 E exceeded precipitation by 25 and 15%, respectively, and in 1999 total E was more than twice the rainfall. At Fyodorovskoye the measurements

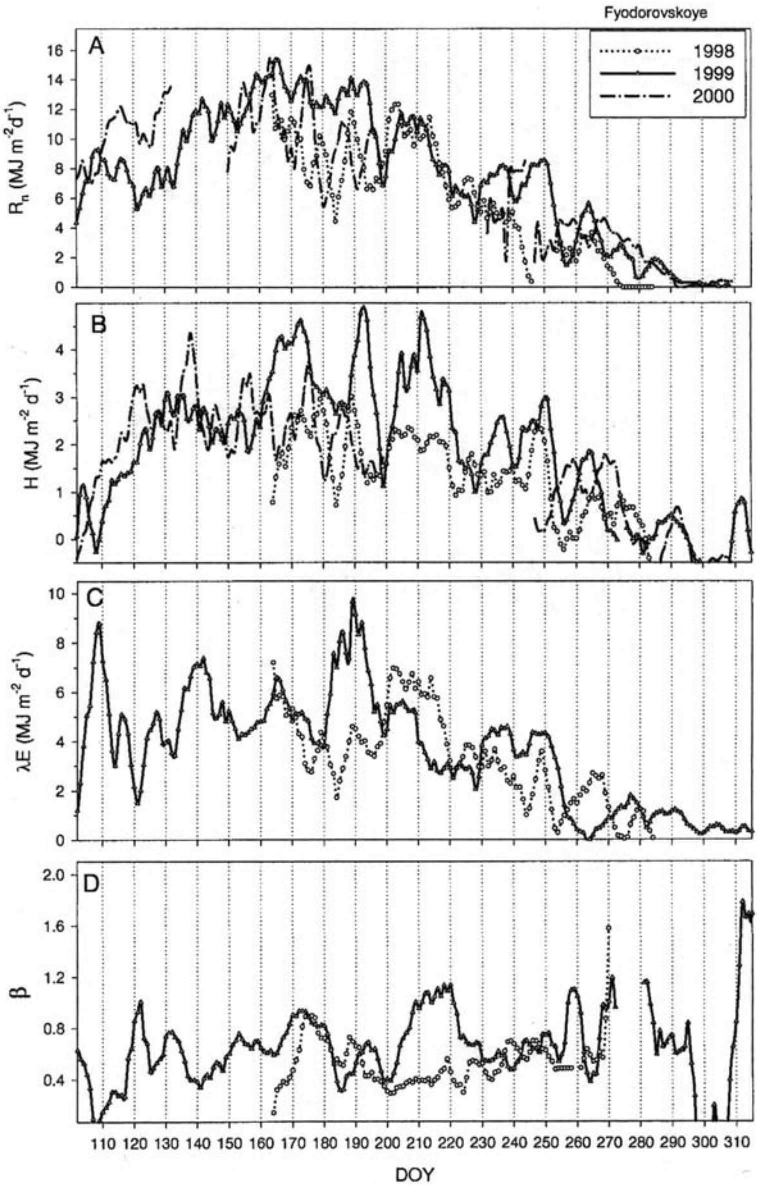

Fig. 5. Five-day running means of daily net radiation, sensible and latent heat fluxes, and Bowen ratio at the Fyodorovskoye site.

covered only in 1999 the entire snowfree period, evaporation in that year totaled 319 mm. Evaporation between 1 June and 30 September was 204 mm, 13% less than at Zotino. On a monthly basis, both R_n and E were lower or similar to what was observed at Zotino. As pointed out by Moncrieff et al. (1996), random errors diminish the more the longer the integration period; over an entire growing season the typical

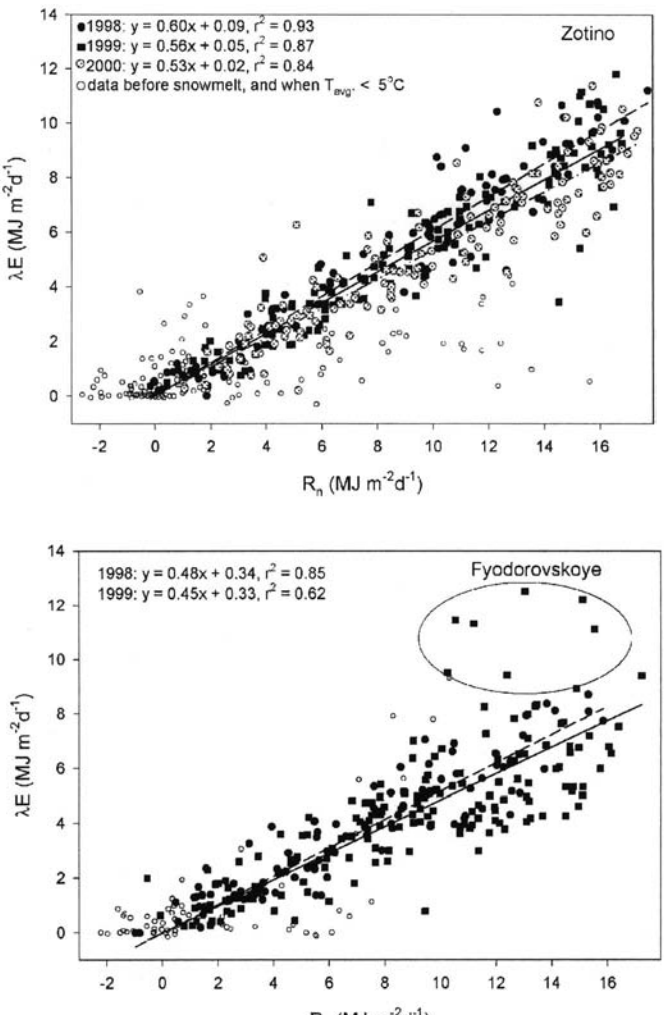

Fig. 6. The relationship between daily net radiation and latent heat flux at Zotino (A) and Fyodorovskoye (B) for 1998 (closed circles), 1999 (closed triangles) and 2000 (grey circles). Data before snowmelt or at days after snowmelt when average temperatures <5 °C are indicated as open circles. The lines are linear regressions through the data where 1998 = dashed line, 1999 = straight line, 2000 = dotted line.

random error magnitude of our instrument setup would be around 10, perhaps 15%. More important with respect to seasonal and annual budgets, however, will be the possible consistent underestimation of energy fluxes, which has been argued for because of the regular lack of energy balance closure in eddy covariance measurements (e.g. Twine et al., 2000). As discussed in section 2.4, precise measurements of all components of the energy balance, particularly of soil heat flux, are difficult to perform in wetlands. We therefore do not know the ratio of $(H + \lambda E)/R_a$ for the two study sites and have not adjusted the fluxes for closure, with the consequence that the numbers given above probably need to be raised by between 10 and 30% (Twine

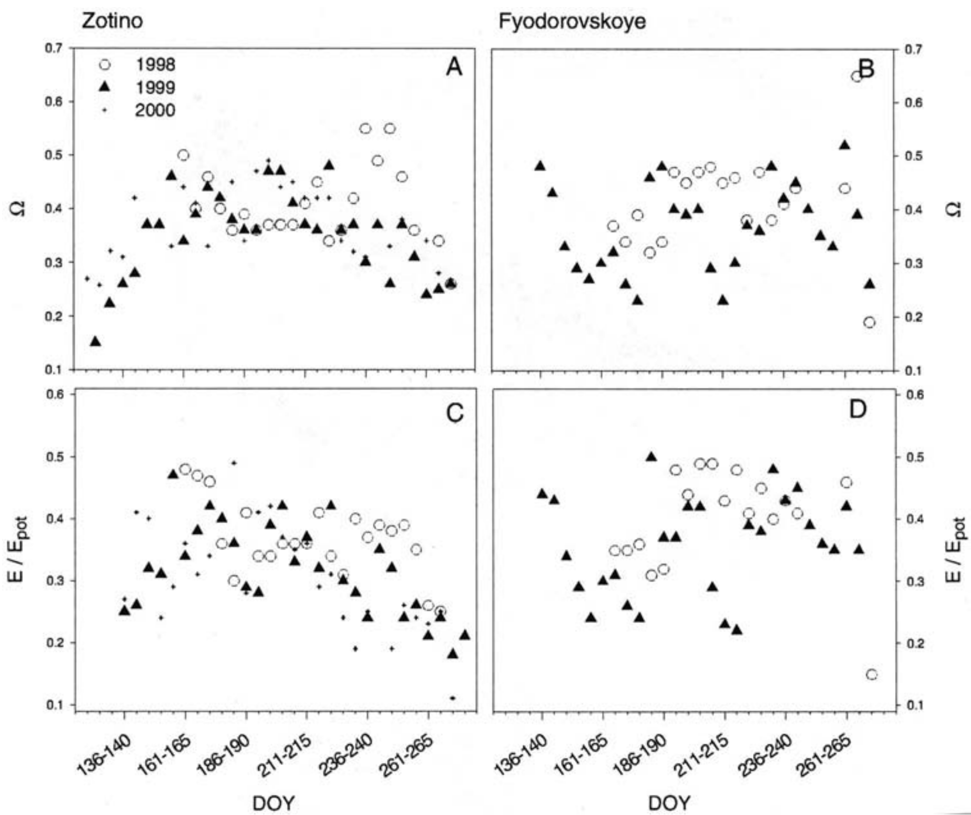

Fig. 7. Five-day average decoupling factors (Ω, A and B), and ratios of evaporation to its potential rate (E/E_{pot}, C and D) at the two study sites. Symbols are given in (A) and are the same throughout the figure. Values are calculated from daylight hours only.

et al., 2000; Tchebakova et al., 2002). This is of importance as the hydrological cycle in our study systems is already tightly closed (see Discussion).

The (de) coupling factor Ω (Jarvis and McNaughton, 1986) separates the contribution of net radiation and atmospheric conditions (windspeed, temperature, vpd) as drivers of total evaporation. Figures 7A and B show 5-day averages of Ω during daylight hours at the two sites. At Zotino, Ω varied between 0.3 and 0.5 for most of the measurement period with a tendency to increase during spring and to decline to values below 0.3 in autumn, except for a ca. 3-week period in September 1998, when Ω exceeded 0.5. Corresponding 5-day average surface conductances (G_s) for water vapour, calculated from the Penman-Monteith equation (e.g. Kelliher et al., 1995) varied between 100 and 400 mmol^{-2} s^{-1} but exceeded 500 mmol^{-2} s^{-1} on single days (not shown). There was no tendency for G_s to decrease progressively with growing season length.

The measured evaporation was compared to its potential rate, calculated by the Penman equation (Monteith and Unsworth, 1990; Lhomme, 1997). At Zotino, the ratio between the two rates varied between average minima of 0.2 and maxima of 0.45-0.5. During spring, the ratio increased steadily, but values in June 1998 were considerably higher than during the same period in 1999 and 2000. In all three years, E/E_{pot} declined relatively rapidly during autumn. During this period, E/E_{pot} in 1998 was higher than in the two years later, but the differences were not as distinct as seen for Ω. or G_s.

At Fyodorovskoye, no clear seasonality in Ω or E/E_{pot} was observed. Ω varied between 0.2 and 0.4 for most of the snowfree period, with the exception of values >0.55 in late July 1998. Overall, values were lower than at Zotino. Missing data for vpd prevented calculation of omega, and E/E_{pot} beyond day 215 in 1998. Ω and E/E_{pot} at Fyodorovskoye varied within a similar range as at the Zotino bog. There was a tendency for E/E_{pot} to increase between spring and midsummer and to decrease in autumn. However, E/E_{pot} during a 3-week period in July/August 1999 were as low as in spring.

4. Discussion

Data from micrometeorological measurements of northern (boreal and subarctic) wetland latent and sensible heat fluxes are scarce, often restricted to measurements that cover few days or a single vegetation period (e.g., Rouse et al., 1987; Lafleur and Rouse, 1988; Lafleur, 1992; den Hartog et al., 1994; Moore et al., 1994; Lafleur et al., 1997; Valentini et al., 2000). The data become even scarcer when restricted to studies in ombrotrophic bogs (den Hartog et al., 1994; Valentini et al., 2000), and our observations thus contribute to elucidate general patterns of boreal wetland evaporation, including seasonal and interannual differences.

Surface albedo between 10 and 20% during the snowfree period, and rapid changes from values between >60% and <20% associated with snowmelt and snowfall (Fig. 1), have also been observed in boreal (Lafleur et al., 1997) and temperate (Kim and Verma, 1996) fens, and are evidently typical for a wide range of wetland ecosystems as moist surfaces are typically relatively dark. In our study, minimum values were associated with periods of maximum wetness directly after snowmelt when most of the ground was still frozen. The generally higher variation in surface albedo measured at Fyodorovskoye during summer (Fig. 1B) reflected the higher and more frequent precipitation (e.g., 1999, doy 220-230, 280; Fig. 2) which was large enough to measurably re-wet the peat surface after dry spells. Albedo in Zotino did not vary significantly on the short-term, but increased gently and continuously until a broad maximum was reached in July. This increase may reflect the gradual lowering of the water table as the growing season progressed as well as a 'greening' of the vegetation as the frost damaged, brown *Sphagnum* mosses developed new chlorophyll after snowmelt and the vascular plants grew new leaves (Lafleur et al., 1997; Nichol et al., 2002).

Average daily evaporation rates during the summer months June-August varied between 1.5 and 3.3 mm d^{-1} at Zotino and between 1.5 and 2.8 mm d^{-1} at Fyodorovskoye. Overall, these rates were within the range previously observed for other wetlands in the northern latitudes. At a boreal fen during a 135-day period between early April and early September, total evaporation was ca. 250 mm or 1.9 mm d^{-1} (Lafleur et al., 1997) while at a subartic fen, evaporation totaled 167 mm in a 60-day period between May and August, 2.8 mm d^{-1} (Moore et al., 1994). During July 1996, average daily E was 2.6 mm d^{-1} at a bog close to the Zotino site (Valentini et al., 2000), slightly less than average rates during all three July months considered here (3.3, 3.0, 2.8 mm d^{-1}). The highest rates measured at Zotino were close to rates measured in coastal marshes in the forest-tundra transition zone in Canada [3.5 mm d^{-1} (Rouse et al., 1987)].

At both study sites, net radiation was the principal factor influencing variation in daily evaporation rates, with no obvious effects due to varying summer precipitation (Fig. 6). The substantial short-term fluctuations in daily λE as well as the significant interannual differences that were observed periodically at both sites could be explained by this single parameter. The proportionality between λE and R_n led to the generally higher summer evaporation at Zotino, where R_n was typically higher (Figs. 3-5). Higher R_n were the result of both higher total downward radiation and somewhat smaller reflectances for total radiation (e.g. 0.75-0.83% at Zotino vs. 0.79-0.85% at Fyodorovskoye; data not shown). Additionally, the amount of radiative energy partitioned into evaporation at Zotino exceeded that in Fyodorovskoye by ca. 10% (Fig. 6). It should be pointed out here that while R_n was the single most significant parameter to explain the observed variation, the absolute magnitude of λE was determined by both radiative and advective terms ($\Omega < 1$, see discussion below).

The dependence of λE on net radiation (or available energy) was also found for a mineral-poor, *Sphagnum*-dominated fen in Canada (Lafleur and Roulet, 1992). However, in northern wetlands with a higher cover by vascular plants, air temperatures (Rouse et al., 1987) and air saturation deficit (Lafleur and Roulet, 1992) were deduced as the dominating environmental variables that constrained λE. Evaporation rates measured in a temperate wetland dominated by *Phragmites* were also independent of R_n (Burba et al., 1999). The available data is limited but suggests a general shift in the regulation of evaporation from radiative control in *Sphagnum* dominated types to control via stomata (and hence via saturation deficit and temperatures) in sedge dominated types.

Overall, the percentage of radiative energy partitioned into evaporating water seems to be highly variable across the range of northern wetlands examined to date. In a subarctic fen 63% of summer R_n were used for evaporation (Moore et al., 1994), while in a boreal bog over permafrost 46% of the available energy was partitioned into E (den Hartog et al., 1994). Bowen ratios in boreal and subarctic wetlands during summer periods are similarly variable: In the boreal bog above permafrost, β was close to 1 during a one-month period (den Hartog et al., 1994), but in three different types of subarctic coastal wetlands β varied between 0.5 and 0.8 (Lafleur and Rouse, 1988). It seems that values in boreal fens can vary between <0.2 and >1 during the summer, with (Lafleur et al., 1997) or without (Moore et al., 1994) a seasonal trend to the variation. During a short-term study in July 1996 at a central Siberian bog located within ca. 30 km of the Zotino site described here, Valentini et al. (2000) found β averaging 0.6. These values indicate that there is some pronounced heterogeneity in energy partitioning even for sites within a relatively close proximity. Variable ratios of ridges, *Sphagnum* lawns and free water surfaces may contribute to this observation. Interestingly at Fyodorovskoye energy partitioning was clearly affected by the frequency and amount of precipitation during the summer months when rainfall and thus re-wetting of the peat surface were detectable as changes in surface albedo (Fig. 8A). Contrasting the observations at Zotino, where no relation between β (and Ω and E/E_{pot}, see discussion below) and albedo was observed, a shift towards energy being partitioned as sensible heat thus corresponded to periods of dry surfaces, characterized by high albedo.

The average values of Ω between 0.2 and 0.5 suggest a relatively strong coupling of the bog's surfaces to atmospheric conditions. In fact, during some periods at Fyodorovskoye Ω were only insignificantly larger than values associated with coniferous forests [0.1-0.2 (Jarvis and McNaughton, 1986)], indicating a strong contribution of advective conditions to total evaporation. At Zotino the ridges with small pine trees growing atop obviously provide enough roughness elements. The even lower Ω at Fyodorovskoye, however, are not as easily explained, as there were basically no trees in a radius of >500 m around the eddy flux

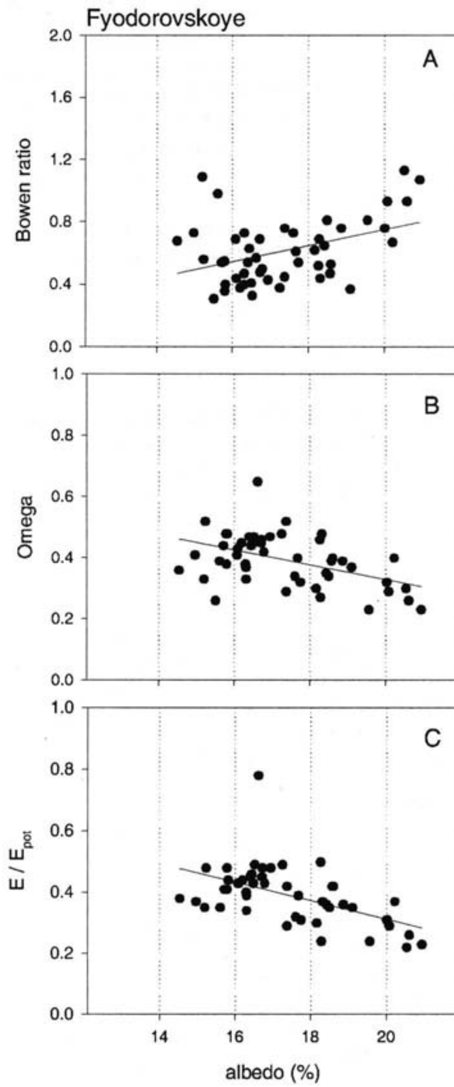

Fig. 8. Five-day averages of Bowen ratio, Ω and E/E_{pot} at Fyodorovskoye as a function of average albedo (a surrogate for surface dryness). Values are calculated from daylight values and restricted to the months June-September 1998 and 1999. Lines are linear regressions through the data.

tower. However, Ω is not solely dependent on surface roughness elements. It will also decrease with surface drying, as then the resistance for heat transfer by convection increases relative to the resistance for water vapour transfer (Jarvis and McNaughton, 1986; Monteith and Unsworth, 1990). Using surface albedo as a surrogate measure of surface dryness we demonstrated that at Fyodorovskoye there was indeed an increase of coupling with the drying peat surface, reflected in the decreasing Ω with increasing albedo (Fig. 8B).

The surface of bog ecosystems is characterised by a significant proportion of free water surfaces in early spring and a significant contribution of non-vascular plants to the total vegetation. The amount of water lost which is controlled by stomatal opening is thus small. This should be reflected in λE being close to a rate observed over a free water surface or by vegetation freely supplied with water. These rates can be calculated as 'equilibrium rates' using the Priestley-Taylor equation (Monteith and Unsworth, 1990) or as 'potential' rates using the Penman equation, respectively, the latter explicitly including the advective influence of air saturation deficit (D) and atmospheric conductance (G_a) in the calculation. It should be kept in mind, however, that the calculated reference evaporation rates are somewhat idealized concepts to investigate the influence of surface dryness. In reality a completely wet surface differs from a dry one in albedo (lower) and temperature (lower), and consequently in R_n and G_a (Lhomme, 1997). Nonetheless, values of $\Omega < 0.5$, as they were evident throughout the study at both bogs, illustrated how equilibrium rates of evaporation clearly were not being attained.

At both sites average E/E_{pot} were always below 0.5. This was perhaps not expected, particularly so in early spring when a relatively large fraction of the bog area was inundated. We used R_n to calculate E_{pot}, which may lead to an overestimation of E_{pot} because of the large soil heat flux in peatlands (i.e. $R_a << R_n$). However, there may also exist additional, more mechanistic explanations for our observations: *Sphagnum* ssp. lack water-conducting tissues and during summer capillary rise is insufficient to supply enough water to keep up with demand (Ingram, 1996). Drying *Sphagnum* can thus act as an excellent 'mulching' substrate, reducing E well below its potential rate as soon as the water table drops on average only few millimetres below the surface (Lafleur and Roulet, 1992). Dead material and plant debris may also protect the free water surface from direct radiation and hence contribute to $E < E_{pot}$ (F. Dunin, personal communication). At Fyodorovskoye summer precipitation was sufficientto restore regularly some of the water lost by evaporation and as for Ω, E/E_{pot} varied with surface dryness as detected by surface albedo changes (Fig. 8C). In contrast and as discussed above, the Zotino rainfall events during summer were not frequent and/or large enough to be measured as changes in short-wave reflectances (i.e. dryness) of the peat surface, hence neither β, Ω or E/E_{pot} was clearly related to changes albedo. However, the seasonal trend of clearly declining E/E_{pot} and Ω from mid-summer onwards may well reflect the progressively declining area of free water surfaces. Correspondingly β increased as relatively more energy was partitioned towards sensible heat flux.

Considering the large differences in precipitation between the two sites, the considerable interannual differences in precipitation and the prolonged periods with little or no rain that were a dominant feature of the climate at both sites, perhaps one of the most surprising results in this study was the lack of influence of growing season precipitation on total evaporation. At Zotino monthly evaporation typically exceeded monthly precipitation, in the driest months by a factor of 5-10; indeed highest monthly rates during the entire measurement period were measured in July 1998, despite only 10 mm of rain. Total summer evaporation in 1999 and 2000 was higher than measured summer rainfall and also higher than the long-term summer averages measured at Bor. In both summers, 47% of long-term annual precipitation evaporated. By contrast, in Fyodorovskoye, monthly evaporation during normally wet months was generally less than precipitation, and only in the unusually dry months June and July 1999 was this pattern reversed. The 1999 summer evaporation was somewhat less than summer precipitation and represented 45% of long-term annual values.

It may be a general phenomenon in northern wetlands that total evaporation during summer can (but not necessarily will) exceed precipitation (Moore et al., 1994; Lafleur et al., 1997; Kelliher et al., 2001). Nonetheless the peat surface does dry out and acts a an efficient mulch that reduces E generally below 50% of E_{pot}. In fact, this mulching effect could be interpreted as 'peat-self-preservation' in the low-precipitation environment. It reduces summer water loss to a sustainable total, as *Sphagnum* ssp. can hold up to 20 times its dry weight as water (Moore, 1997). An important feature of northern wetland hydrology thus is the spring recharge of the peat water storage from melting winter precipitation. Were evaporation running closer to its potential rate then peat water storage would not be sufficient to keep up with demand during the warm and dry summers. As has been discussed for boreal forests (Kelliher et al., 2001), the hydrological cycle in boreal wetlands is thus tightly closed and possibly rather vulnerable to changes in annual precipitation.

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