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Net ecosystem exchange of carbondioxide and water of far eastern Siberian Larch (*Larix dahurica*) on permafrost

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Abstract

Observations of the net ecosystem exchange of water and CO₂ were made during two seasons in 2000 and 2001 above a Larch forest in Far East Siberia (Yakutsk). The measurements were obtained by eddy correlation. There is a very sharply pronounced growing season of 100 days when the forest is leaved. Maximum daytime uptake rates are 18 μmol m⁻² s⁻¹; maximum respiration rates are 5 μmol m⁻² s⁻¹. Net annual sequestration of carbon (C) was estimated at 1.7 (±0.5) ton C ha⁻¹ in 2001. The net carbon exchange of the forest was extremely sensitive to small changes in weather that may switch the forest easily from a sink to a source, even in summer. June was the month with highest uptake in 2001.

The average evaporation rate of the forest approached 1.46 mm day⁻¹ during the growing season, with peak values of 3 mm day⁻¹ with an estimated annual evaporation of 213 mm, closely approaching the average annual rainfall amount. 2001 was a drier year than 2000 and this is reflected in the evaporation rates that show lower evaporation rates in 2001 than in 2000.

The surface conductance of the forest shows a marked response to increasing atmospheric humidity deficits. This affects the CO₂ uptake and evaporation in a different manner, with the CO₂ uptake being more affected. There appears to be no change in the relation between surface conductance and normalized net ecosystem uptake at the monthly time scale. The response to atmospheric humidity deficits is an efficient mechanism to prevent severe water loss during the short intense growing season. The associated cost to the sequestration of carbon may be another explanation for the slow growth of these forests in this environment.

1. Introduction

There is increasing evidence that the northern latitudes are experiencing the effects of global warming. The increased and early greening of the land surface as detected

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by satellite remote sensing and forest inventory data (Buermann et al., 2003) is one of the strong lines of evidence. There is also evidence, both from inverse and bottom up modelling studies (e.g. Bousquet et al., 1999; Lucht et al., 2002) that the northern hemisphere is sequestering large amounts of carbon at increased rates. It is thus of considerable interest to determine the carbon uptake of forest in the Northern hemisphere. Although Siberian forests constitute 20% of the world's forest area, little is known about their role in the carbon budget and about their role in the regional and continental water balance.

Preliminary studies using eddy correlation estimates, mostly from Central Siberia, indicate that the sink strength of Siberian pine forest is between 0.5 and 2.5 ton C ha⁻¹ yr⁻¹ (Schulze et al., 1999). Estimates using these atmospheric modelling techniques suggest a carbon sink capacity of 1.5 Gton C yr⁻¹ for North Asia (Bousquet et al., 1999). The latter estimate includes all land use change in a ten-year period and is based on atmospheric CO₂ measurements. More recently Roedenbeck et al. (2003) suggest an approximately neutral carbon balance for boreal Eurasia. These results obtained using inverse modelling techniques are poorly constrained by the observations and, clearly, more information on the carbon balance of boreal Eurasian forests is needed to better define the a-priori estimates that are used in the inversion studies.

The forests of Siberia represent one of the last natural frontiers in the world. Nearly 65% of these forests grow in areas with permafrost (Shvidenko and Nilsson, 1994). The Siberian forests in the Far East cover 45% of the total forests in Siberia. It is estimated that 74 Gton C and 249 Gton C is stored in the vegetation and soil respectively, of forest ecosystems of Siberia (Dixon et al., 1994). The estimated carbon stocks in the soils of forest and tundra ecosystems of Yakutia (East Siberia) amount to 17 Gton under the forests cover (in total 125.5 Mha forest and 37 Mha tundra). This is about 25% of the total carbon stocks of forest soils in the Russian Federation.

There has been some scattered previous work in this area of Siberia that, amongst others, stimulated the current study. Kelliher et al. (1997) took a series of 9 days of eddy correlation measurements above a forest some 160 km South of Yakutsk and

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found average daily evaporation to be at 1.9 mm (± 0.3). Ohta et al. (2001) took a full year of measurements and maximum rates of 2.9 mm day⁻¹ at the beginning of July. The average rate over the growing season was estimated at 1.2 mm day⁻¹. Ohta et al. (2001) also found a strong seasonality in the evaporation fluxes that was, not unexpectedly, related to the existence of needles on the canopy. Based on their respective measurements and some interpolation, Kelliher et al. (1997) and Ohta et al. (2001) estimated total evaporation of East Siberian Larch to be at 169 and 151 mm, respectively. With an annual precipitation of 213 mm, this leaves preciously little water available for runoff.

For Central Siberian forest Schulze et al. (1999) in an extensive review, quote daily evaporation rates for a larch and two pine forests of 1.4 to 1.7 mm in July that appear to be somewhat lower than those for East Siberian larch. An analysis of two years of measurements over a Central Siberian pine forest by Tchebakova et al. (2002) found evaporation rates of 1.5 to 2 mm day⁻¹ with a three year average for the growing season of 1.5 mm day⁻¹. Generally these low evaporation rates, typically using up to only 20% of the available energy, are associated with high Bowen ratios, well above 1, even for a forest without any substantial water stress.

Maximum rates of Net Ecosystem Exchange of CO₂ (NEE) quoted by Röser et al. (2002) for West Siberian Betula and two mixed stands are 13, 10 and 8 $\mu\text{mol C m}^{-2}\text{s}^{-1}$, respectively. For the same forest as discussed in this paper Hiyama et al. (2001) quotes maximum uptake rates for July of 15.9 $\mu\text{mol C m}^{-2}\text{s}^{-1}$. Schulze et al. (1999) quote maximum daytime values in July between 7 and 11 $\mu\text{mol C m}^{-2}\text{s}^{-1}$, again for Central Siberian forests.

At present there is little empirical understanding of the role of this stock in the global carbon cycle, or, perhaps more importantly, how it may change in the future under changing natural (fire, climate) or anthropogenic forcing (logging). It is well known (e.g. Lindroth et al., 1998) that European boreal forest on drained peat soils can be large sources rather than sinks of CO₂ in years when early thawing sets in. If global warming in the boreal forest region of Eurasia becomes persistent, as suggested by

Serreze et al. (2000), then the Eurasian forest on permafrost may experience a similar shift from an arguably small sink to a much larger source. The sheer magnitude of the area involved makes this an important issue for research. The key to understanding this behaviour is to investigate the sensitivity of Net Ecosystem Exchange (the balance between assimilation, and heterotrophic and autotrophic respiration, NEE) in situ. Ideally this would be augmented by longer-term estimates of disturbance such as fires, to assess the Net Biome Production (NBP) (e.g. Körner, 2003; Dolman et al., 2003), but a process understanding at annual timescales is an obvious prerequisite.

Thus, to be able to give reliable estimates of carbon sequestration of Far East Siberian forests direct measurements of the net uptake of CO₂ at seasonal to annual timescales are required. An advantage of such direct measurements is that they also give insight into the sensitivity of the eco-physiology of Siberian forest to changes in climate. In this paper eddy correlation measurements of CO₂ flux and evaporation over an undisturbed Larch (*Larix dahurica*) forest near Yakutsk are presented. This part of the forest may be considered representative of the vast expanse of larch forest of East Siberia.

The little information available today about Siberian forest relates primarily to Central Siberia or concerns relatively short periods of campaign based measurements (see also Heimann, 2002). This paper aims to extend that information and describes direct measurements of net ecosystem exchange of CO₂ and water and energy fluxes of a larch forest in East Siberia, near Yakutsk. It first describes the site and instrumentation, and then discusses results from the measurements campaign. It presents the first series of coupled evaporation and CO₂ exchange observations over a two-year period. This allows also within specific uncertainty ranges, the annual sink strength to be determined.

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2. Site description and methods

2.1. Site description

The forests around Yakutsk form part of the vast watershed of the River Lena with an estimated surface area of 2490 km². The territory of the Sakha Republic (Yakutia) covers some 3103 thousand km², including the New Siberian Islands. The largest rivers are the Lena, Anabar, Olenek, Yana, Indigirka and the Kolyma. Most of this territory is covered by forest; toward the north the forest changes into tundra lands. A useful, up to date review of the geology, climate and ecology may be found in Giorgiadi and Fukushima (1999).

The measurement site is located in the middle reaches of the Lena and is in a region of continuous permafrost. The climate exhibits a strong continentality; at Yakutsk the annual mean temperature is -10.4°C and the absolute minimum is -57.1°C . The mean annual rainfall in Yakutsk is 213 mm (e.g. Schulze et al., 1999); the 30-year climatological average is 240 mm. The soils in the area consist of fluvial deposits and are classified as cryomorphic derno-taiga solidized soils. Low annual mean rainfall prevents podzolization in this area. Crucial is the influence of the permafrost layer, which thaws down to 1.2 m below the forest floor during the summer and then freezes up again during the autumn and winter. The dominant species in the forests is Dahurica Larch: *Larix dahurica*. The forests are best classified as “middle taiga” or light taiga.

The forest where the measurements were taken lies about 40 km northeast of the city of Yakutsk, at the Forest station “Spasskaya Pad” ($62^{\circ}15'18.4''\text{N}$, $129^{\circ}37'07.9''\text{E}$). The altitude is 220 m a.s.l. The site is described extensively in Ohta et al. (2001). At the time of measurement the mean stand height was 18 meter and the stand density 840 trees ha⁻¹. The average age of the stand is 160 years. The Japanese-Russian team of Ohta et al. (2001) established a scaffolding tower in 1996 of 32 m high, which was used to obtain the current set of measurements. In 2001 no separate estimates of plant area were available, but the estimates of Ohta et al. (2001) for the leafless canopy of 1.7 and 3.7 for the fully leaved season, may serve as a useful reference. It

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is relevant to note that due to local fire protections, there has not been fire at the site over the last 80 years.

2.2. Methods

An eddy correlation instrument consisting of a 3-D Gill Solent sonic anemometer (R2), a krypton hygrometer and a LICOR 6262 infrared gas analyser were installed at 34 m on a telescopic mast, mounted on the micrometeorological tower. This system measured the net ecosystem flux (NEE) and also latent (evaporation), sensible heat and momentum exchange (e.g. Aubinet et al., 2000). The covariances as well as the raw data were stored and post processed using software that corrects for sensor misalignment, frequency loss, sensor separation etc. (Aubinet et al., 2000; Dolman et al., 2002). We also included the flux angle correction as proposed by Gash and Dolman (2003) and van der Molen et al. (2004).

The measurements of NEE were taken over a two-year period, from 14 July 2000 until 1 December 2000 and the next year from 20 April 2001 to 25 September 2001. Figure 1 shows the degree of energy balance closure for the daily 2001 data. A regression line gives a slope of 0.88 with an intercept of 26 and an r^2 of 0.79 (a regression forced through the origin gives a slope of 0.92). The addition of the flux angle correction (van der Molen et al., 2004) improved the energy balance closure by 13%. This degree of energy balance closure gives good confidence in the quality of our flux measurements. We suspect that the remaining energy loss is due to a mismatch of the footprint of the radiative sensors and the eddy correlation instruments. Spectra (not shown) show good agreement with the classic Kaimal shapes (Kaimal and Finnigan, 1994).

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3. Results

3.1. Net ecosystem exchange of CO₂

Figure 2 gives the seasonal and daily course of the half-hourly eddy correlation measurements for the CO₂ flux plotted as contour lines. We show here both the 2001 set as this covers a full seasonal cycle, and the 2000 dataset that started only in July. We concentrate first on the 2001 measurements. From the start of the 2001 measurements (20 April 2001, day 110) until day 144 the forest loses a small amount of carbon by soil respiration (positive NEE). Around day 130 (10 May) there is a sudden increase in respiration. This appears to be primarily related to increased temperatures, stimulating heterotrophic respiration, that drop down again after that day. Interestingly there is also a small peak around day 140 (20 May) when the NEE is positive. We suggest that some of this peak may be related to pre-budding autotrophic respiration, caused by the trees starting to produce the needles. From day 144 the forest starts gaining carbon and in only about 3 weeks (day 165) the forest is taking up carbon at maximum rates of about $18 \mu\text{mol m}^{-2}\text{s}^{-1}$. After mid-July, around day 200, the uptake decreases in a particularly strong fashion in the afternoon, but recovers somewhat around early August (day 220). After that, there is an almost steady decrease down from about $10 \mu\text{mol m}^{-2}\text{s}^{-1}$ towards $0 \mu\text{mol m}^{-2}\text{s}^{-1}$ at the end of the season.

Unfortunately we have no measurements in June 2000 due to technical and custom problems that would allow a full seasonal comparison with the 2001 data. However, there is similarity in the early August uptake but difference in the autumnal uptake of 2000 and 2001. The difference is that NEE in 2000 remains slightly longer negative adding a few extra days of uptake to the annual total. Furthermore the peak values of uptake are still at $15 \mu\text{mol m}^{-2}\text{s}^{-1}$ while those in 2001 are approaching only 60% of that value. The similarity between years, however, is remarkable in the sense that in both 2000 and 2001 there appears to be a dip in the uptake around late July, after day 200. The timing of this decline in uptake remains noteworthy, and we will come back to this later in the paper. We cannot, of course, rule out the possibility that the observed

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similarity in timing is fortuitous, and the result a completely random phenomenon, but we believe there that there may be a realistic interpretation possible. In 2001 there is total of 100 days of negative (uptake) CO₂ flux from 26 May to 7 September (days 146–251).

5 The average diurnal trend for the CO₂ flux can also be inferred from Fig. 2. It is evident that only three months contribute to the net seasonal uptake: June, July and August. The highest uptake takes place in June 2001. Comparing the colour and shape of the July 2000 and 2001 curves, it appears that in 2000 the uptake is similar at peak times (about 18 μmol m⁻²s⁻¹). The July curves are similar up to noon, but
10 after that, NEE starts to drop off sharply in 2001 in comparison to those of 2000. This leads to a lower uptake in July 2001 than in July 2000 (see also Table 1). August 2001 (day 210–240) also shows a substantially lower rate than in 2000, with the previously noted similarity in timing of the decline. The uptake is almost finishing at 15:00 h. We hypothesize that reductions like this are related to the closing of stomata at high vapour pressure deficits, whereas overall reductions, that show less diurnal variation
15 are caused by soil moisture deficits that occurred in 2001, but not in 2000. This will be discussed later down in the paper. In Fig. 2 these effects show as a gradual weakening from top to bottom following the daily cycle and as a weakening of the high uptake colours from right to the left, as the season progresses.

20 The other effect that can be observed is the lengthening and shortening of the daylight period during the season. In June the forest uptake starts at 06:00 h. and continues to 22:00 h at night, because the night is only 3–4 h long during June at this latitude.

As suggested by the similarity of the daily patterns in both years, the seasonal uptake also exhibits a clear pattern. This is shown as the mean monthly NEE values in Table 1. On average, the forest loses carbon in April and May, while the uptake is strongest
25 in June (at least in 2001), followed by a small decline in uptake in late July with a further decline in August. In September the larch has shed their needles, or photosynthetic activity has ceased, and until the soil gets frozen, the soil and trees continue to respire and loose carbon. In 2000, NEE in July is comparable to the July 2001 values. The

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overall rates for August and September are considerably lower in 2001 than in 2000.

Unfortunately we were not able to obtain reliable estimates of the storage terms that could be used to correct the eddy correlation measurements for storage to obtain the biotic flux. The forest is relatively sparse, so we expect that our daytime estimates are not substantially affected by this omission. Furthermore, the larch site has little topography that would influence or generate strong drainage flows at night, that would dramatically impact our night-time fluxes. To investigate the impact of storage, we applied two types of corrections to our night-time data that compensate for the lack of storage measurements. Both methods assume that at high friction velocities the eddy correlation system gives “true” estimates of the flux and that the storage is negligible under these conditions. The first consists of the “standard” replacement of fluxes obtained at low friction velocities by values that are obtained from a curve relating temperature and night-time flux fitted for values of high friction velocity ($u_* < 0.6$). The second assumes a linear relationship between the observed flux and that at a threshold friction velocity (0.6 m s^{-1}) and scales the eddy correlation values at low friction velocities to that threshold value. This is equivalent to assuming that the “missing” flux is compensated for by an ideal storage flux, as would be common practice at sites with no night-time problems (Aubinet et al., 2000). Table 1 gives the average monthly fluxes obtained by these two methods. Although an independent estimate of the storage flux would be preferable, we believe that by using these two methods we can provide an indication of the uncertainty related to the storage. There is still considerable uncertainty attached to our night-time estimates. This uncertainty also makes a comparison between the night-time fluxes of the last three months of the growing season between 2000 and 2001 rather ineffectual. Leaving out the u_* correction, would increase our estimated annual uptake from 12.1–12.7 in g C m^{-2} 2001 to 17.7 g C m^{-2} . So, clearly, despite the absence of topography, a correction for the night-time fluxes is appropriate.

Figure 3 shows the differences in total, daytime and night-time NEE in more detail. When looking at day-time and night-time fluxes separately, it becomes clear that the high rate in June 2001 is primarily caused by high day-time uptake. This is a consistent

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feature as June is the month with the highest physiological activity of this system. In contrast, the highest night-time fluxes are obtained at the warmer months of July and August.

3.2. Evaporation and energy balance

5 The monthly average evaporation rates are given for 2000 and 2001 in Table 2. The highest evaporation rates for 2000 occur in July (no measurements in June, however) and those of 2001 in June, with little difference between June and July. Evaporation rates in 2000 are consistently higher than in 2001. It is worth noting that the eddy correlation system “sees” both soil and plant evaporation, and there may be some contamination as a result of evaporation from a wet canopy immediately after rainfall. In general however, the measurements system does not yield reliable values during rainfall, so those data were ignored. Overall, the evaporation results thus relate primarily to transpiration with a small contribution of soil evaporation on days immediately following rainfall event. We were not able to measure soil evaporation directly. The under storey contribution was also not measured separately, but Ohta et al., suggest that this may contribute to 35% of the total evaporation. Given the denseness of the under storey we neglect soil evaporation, and treat the evaporation from the dry understorey and upper storey as a single entity.

Figure 4 shows the diurnal and seasonal course of evaporation (latent heat flux) for 2000 and 2001. In 2001, evaporation is almost zero at the start of the growing season and quickly increases to values of 75 Wm^{-2} later in the season in 2001. Much higher rates were observed in 2000. The rapid increase in latent heat flux coincides with the rapid increase in photosynthetic activity observed in Fig. 2 and is related to the onset of needle growth

25 Taking the values of August (day 210–240) as a reference we note that evaporation rates in 2000 are substantially higher than in 2001. From the end of July 2001 onwards a rather sharp decline in evaporation can be observed. In contrast, evaporation during the autumn of 2000 continues to decline at a much more gradual rate. This suggests

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2001 that either physiological or climatic control on evaporation was more strongly present in 2001 than in 2000. We will explore this issue in a later section in this paper. In support of this, Sugimoto et al. (2003) note that soil moisture content in 2000 was much higher than the previous years, causing high evaporation and transpiration rates, or perhaps more precisely, the presence of enough soil moisture caused evaporation and NEE to be non-restricted.

To indicate how much soil moisture stress the forest is experiencing, we calculated a running water balance by subtracting the daily measured evaporation from the observed rainfall at Yakutsk airport. This is shown in Fig. 5. The insert shows the monthly rainfall distribution for 2000 and 2001. It is clear from this graph that in the growing season only July generates a large difference in rainfall, with July 2001 the driest month. The largest rainfall deficit approaches 60 mm, towards the end of the growing season. It is worth noting that at this time, the thawing depth approaches 120–140 cm (Sugimoto et al., 2003). For the loamy sandy soil around the site, Sugimoto et al. estimate the total soil water content at roughly 400 mm, with some 10% of that supplied by spring snow melting in 2000. However, the variation in soil moisture content below 60 cm, roughly half of the thawing depth is small. Once the clay soil is saturated after melting, the top 30–40 cm soil would contain roughly 100 mm of water (Sugimoto, et al., 2003). Towards the end of July this water would be depleted by a maximum of 60 mm according to Fig. 5 and the remaining moisture would become harder to extract. In both 2000 and 2001 soil moisture was approaching this limit, but soil moisture was probably not yet severely restricting evaporation and photosynthesis.

Sensible heat flux is the larger term in the energy balance during most of the growing seasons in 2000 and 2001. Bowen ratios (the ratio of sensible heat and latent heat) are generally well above 1. In 2001, from days 150 to 250, the Bowen ratio

β , is at its lowest value, with values between 1 and 2. When the forest becomes less active, the Bowen ratio quickly rises to values above 10. There is a sharp increase in evaporation after day 150 and a sharp stop at day 250, related to needle growth and senescence, but as can be seen from Figs. 2 and 3, it is certainly likely that thawing

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of the top soil layer at the start of the season initiates root activity and that this activity subsequently triggers the growing of the needles. The evaporation early in the growing season is thus probably related to evaporation of melted snow and soil moisture rather than evaporation from the still needle-less canopy.

5 Figure 4 also shows that evaporation in the summer months of July and August 2001 is repressed compared to that in the same period in 2000. Evaporation in August 2001 is almost a third of that observed in August 2000. This suggests a severe restriction on evaporation. In June and July 2001 the diurnal trends in evaporation are much more similar than for NEE (Figs. 2 and 4). The observed afternoon drop in evaporation
10 for July 2001 appears smaller than for NEE. This requires the investigation of possible causes of this drop and an explanation why NEE and evaporation appear to have different responses to limiting environmental factors.

3.3. Surface conductance, vapour pressure deficit and the control of NEE and evaporation

15 The surface conductance, g_s , is plotted against specific humidity deficit δq in Fig. 6b. The surface conductance was obtained by inverting the Penman-Monteith equation allowing for differences in heat and momentum transport and stability (Dolman et al., 2002). There is sharp response of the conductance to a change in specific humidity deficit. At high values ($\delta q > 15 \text{ g kg}^{-1}$) the surface conductance is only 30% of the value
20 at low values. Note that this data includes data at various levels of solar radiation. No attempt is made to separate this into classes of radiation, as there was virtually no difference in the relation obtained, when the data was grouped in four classes of radiation each covering 250 Wm^{-2} (not shown).

In Fig. 7 the diurnal and seasonal course of the specific humidity deficit is plotted for
25 the two years of measurement. Sharp peaks in the specific humidity deficit appear late in the afternoon and are particularly pronounced in both years around days 200–210 (late July). This coincides with the period during which we also observed a dramatic decline in NEE (Fig. 2). In July 2001 the specific humidity deficit is substantially larger

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than in 2000 (13 g kg^{-1} in 2001 versus 7 g kg^{-1}). According to Fig. 6b this would imply an average value of g_s of about 2 mm s^{-1} in 2000 and 3 mm s^{-1} in 2001. This corresponds remarkably well with the observed values. Note that the maximum values of the conductances could be larger, but the binning procedure applied to produce Fig. 6 reduces the overall magnitude.

The reduction in conductance appears to have different effects on the CO_2 and evaporation flux. The high specific humidity deficit in July 20001 leads to a sharp drop in CO_2 uptake (Fig. 3), but less so in the evaporation flux, because, although the conductance is reduced by almost 50%, the atmospheric demand has also increased. In Fig. 6a the response of evaporation is plotted against surface conductance and saturation can be observed to occur after $g_s=6 \text{ mm s}^{-1}$. In aerodynamically rough vegetation, the evaporation can be approximated by $\lambda E = (\rho C_p / \gamma) g_s \times \delta q$. This relationship leads to a saturation curve of evaporation for a wide range of specific humidity values (Fig. 6d), as can be understood when a simple negative linear relation between conductance and specific humidity deficit is inserted into the simplified equation for evaporation ($g_s = g_{\text{smax}} - a \delta q$ where g_{smax} is the maximum conductance and a a fitting parameter). A reduction in conductance with a corresponding increase in specific humidity deficit does then not lead automatically to a direct change in evaporation, whereas a reduction in g_s immediately causes a decrease in CO_2 uptake (Fig. 6d). If we select the specific humidity deficit values from 10 to 20 g kg^{-1} , Fig. 6d shows that there is minimal variation in evaporation over this range. This explains why, despite a considerable variation in specific humidity deficit, June and July 2001 have nearly similar evaporation rates.

For the same range of specific humidity deficits and corresponding conductances, however, NEE drops from an average value of $4 \mu\text{mol m}^{-2} \text{ s}^{-1}$ to almost zero. A further interesting phenomenon is that the peak of NEE and evaporation when plotted against specific humidity deficit are shifted with respect to each other. NEE reaches an optimum at lower specific humidity deficit than evaporation. This would be of obvious ecological benefit for a forest with a short growing season, of which a considerable part

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towards the dry end of the season exhibits very dry atmospheric conditions, that would very rapidly deplete the scarce soil moisture resource in the top soil where most of the active roots are, if such a strong feedback would be absent.

It appears that the higher evaporation values in the summer of 2000 may be explained by the absence of significant atmospheric constraints on the conductance through vapour pressure deficit. The work of Sugimoto et al. (2003) at the same site suggests that soil moisture deficits may not be that restrictive in 2000 as melting of lower permafrost water provides a continuous source of water to the trees (see also above, Sect. 3.2). In 2001 we had no observations of soil moisture.

Lloyd et al. (2002) found the somewhat surprising result that the water use efficiency of central Siberian pine, expressed as their parameter γ , describing the change of evaporation over a change of photosynthesis, was different for their two years of analysis. In Fig. 8 we plot a similar quantity, but in line with Dolman et al. (2002) we plot the canopy level conductance as a function of net ecosystem exchange as expressed by the Ball et al. (1987) linear relationship between these variables once photosynthesis is normalised by internal CO_2 concentration and relative humidity. We plot monthly values and although the scatter, particularly towards the high conductance end, is rather large, there appears to be little difference between the growing season months in 2000 and 2001. Particularly values at the high conductance end of the curve may be contaminated by soil evaporation. August 2000 shows a somewhat higher slope, but given the uncertainty associated with the linear regressions, we cannot conclude that this slope is different from the others. The normalised water use efficiency may thus be a fairly constant characteristic of this larch system and not vary substantially between months and years. Variations in the actual water use efficiency (not the normalised values of Fig. 8, that show no effect of atmospheric humidity) occur largely as a result of changing environmental conditions, such as vapour pressure deficits during summer. In fact, the similarity between these relations for several months when relative humidity is used to normalize gives additional evidence for the importance of the specific humidity deficit response on conductance as discussed earlier.

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3.4. Annual carbon sequestration and evaporation of far eastern Siberian Larch

Total NEE during the growing season, obtained by summing up daily values, is estimated as 12.1 or 12.7 mol C m⁻² (145 or 152.4 g C m⁻²) depending on which night-time correction method we use. Our two correction methods, once we apply the correction for friction velocity, do not have a significant effect of the annual NEE, which is reassuring. In the appendix we develop an approximation for the respiration that occurs when the permafrost starts freezing up again, from both the top and below. This leads to a small extra amount of respiration of 1.6 mol C m⁻² (19.2 g C m⁻²). Thus we estimate the total CO₂ uptake of this forest in 2001 to be 164.2–171.6 g C m⁻² C (~1.7 ton C ha⁻¹). This assumes, probably correctly that there is no loss in the period January–March because of the frozen condition of the soil. Hiyama et al. (2001) estimated a somewhat higher uptake of 210 g C ha⁻¹ yr⁻¹, but did not take into account the extra loss due to CO₂ trapped in the soil. Because the peak uptake occurs in June, it appears not reasonable to try to expand our 2000 measurements to a full year. Given the somewhat higher rates in August and September 2000, it is however to be expected that the annual NEE would have been slightly higher in 2000 than in 2001.

The average dry canopy evaporation (transpiration) rate in 2001 is 1.46 mm day⁻¹ for the growing season. Peak values, in early July are up to 3 mm day⁻¹, while in September and early April the rates approach zero. Our rates compare well with the 1.25 mm day⁻¹ found for Scots Pine in the Middle Siberian taiga (Tchebakova et al., 2002) and 1.49 mm day⁻¹ found by Ohta et al. (2001) for the same forest in Yakutsk. Based on a 100-day growing season this would put the annual evaporation at 146 mm year⁻¹ and close to the total value estimated by Ohta et al. (2001), of 151 mm. Our values may include some wet canopy evaporation (interception evaporation), however, as there is little rainfall during summer, and data collection in general is poor during rainfall, the impact of including this small, fraction of wet canopy evaporation in our estimates of dry canopy evaporation will be small.

A 100 day growing season may be regarded as the minimum period, based primarily

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on our CO₂ measurements. Expanding it realistically by 20 days (see Fig. 4) would bring the total dry canopy evaporation close to 175 mm. Ohta et al. (2001) suggested that 15% of their total evaporation consisted of evaporation of intercepted rainfall. The addition of 32 mm interception loss would bring the total evaporation close to 200 mm per year. This is very close to the mean annual rainfall of 213 mm, suggesting there would be little precipitation available for runoff.

4. Discussion and conclusions

A first annual estimate of the uptake of East Siberian larch is estimated to be 1.7 ton C ha⁻¹ yr⁻¹, based on our measurements in 2001. Note that this is not the long-term carbon sequestration that is called Net Biome Production. Our estimate is best regarded as Net Ecosystem Production, NEP (Körner, 2003; Dolman et al., 2003). To arrive at the long-term sequestration (NBP), the effects of natural disturbances such as fires would have to be accounted for. Forest fires, harvest and other disturbances will lower NEP substantially. In the longer term it is likely that about 10% of this net ecosystem sequestration gets locked up in soil carbon as Net Biome Production. According to Shvidenko and Nilson (1994) the total forested area of larch in the Far East of Siberia is 169·10⁹ m². This would yield an annual sequestration capacity (NEP) of larch in Far East Siberia of 280 Mton C. This is likely to be an upper estimate and may vary considerable from year to year.

Hollinger et al. (1998) took 14 days of measurements over a larch forest south of Yakutsk and reported very low maximum rates of midday net ecosystem exchange. The values we report here are substantially (a factor 2) higher. Both Hollinger et al. and our work use eddy correlation equipment and obtained good energy balance closure. Assuming that our measurements are correct, there are three possible explanations for this discrepancy. One relates to the possibility that the values obtained from Hollinger et al.'s series of only 14 days just randomly happened to be on the low side. We cannot completely rule out this possibility, and our year-long measurements suggest that after

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the peak in June rather sharp declines in uptake may be observed. It would appear however that these are not sufficient to explain the large difference in maximum flux. The second explanation, and to some extent more worrying one, would hold that large spatial differences due to topography and local site conditions may cause such large differences. The fact that these forests occur on permafrost areas where thawing and thermokarst processes may provide rather large differences in site conditions could potentially play a role. If that is the case, the use of single site measurements to represent large areas in global models is even more dangerous than previously assumed. A similar and related explanation would take into account age and density differences in the forest. The Spasskaya Pad forest is in fact older than the Hollinger et al. (1998) forest: 160 versus 125 year. It appears unlikely that this age difference would explain the current differences: one would expect an older forest to take up less carbon. However there is considerable difference in tree density (1000 versus 840 trees per hectare) that may be important. If this were the main cause, the implication for large scale modelling would be that density distribution as related to age and fire frequency would be crucial information.

The results obtained in this study can be compared with other studies on Siberian ecosystems. The overall rates of NEE compare well with those given by Röser et al. (2002) and Lloyd et al. (2002) for Central Siberian forests, but are somewhat higher. The most noticeable phenomenon in the present case is the very sharp increase in negative NEE (uptake) after initial needle growth. The forest is completely dormant for nine months of the year and suddenly switches to photosynthesis in June. June clearly is the month with the highest uptake, whereas for the Central Siberian forest this appears to be July (Röser et al., 2002).

Our summer rates are high ($18 \mu\text{mol m}^{-2}\text{s}^{-1}$): maximum rates quoted by Röser et al. (2001) for West Siberian *Betula* and two mixed stands are 13, 10 and $8 \mu\text{mol m}^{-2}\text{s}^{-1}$, respectively. Lloyd et al. (2001) quote maximum median values of $12.5 \mu\text{mol m}^{-2}\text{s}^{-1}$, but maximum rates are more in agreement with those of us. In the case of the current larch forest Hiyama et al. (2001) quotes maximum uptake rates for July

of $15.9 \mu\text{mol m}^{-2}\text{s}^{-1}$, comparable to the current values. The high daytime rates are further substantiated by independent measurements of photosynthesis in 2003 that show maximum rates at leaf level in July of about $12 \mu\text{mol m}^{-2}\text{s}^{-1}$. With a leaf area index of around 2 this would yield maximum photosynthesis rate at canopy level of $24 \mu\text{mol m}^{-2}\text{s}^{-1}$. This adds credibility to the present estimates, and suggests that CO_2 uptake rates of larch forest in East Siberia at peak times may be higher than those of pine forest in Central Siberia.

The evaporation values are somewhat higher than previously quoted. The only data that are comparable to those presented in this paper are those of Kelliher et al. (1997), Ohta et al. (2001), Kelliher et al. (1997) only presenting measurements only for a few days in June. These authors suggest that there is little control of surface conductance on evaporation and CO_2 exchange. Apart from the fact that theirs is a different site and the results pertain only to a very short period of June 1993, they also show a substantial reduction in conductance and transpiration as a result of increasing vapour pressure deficit. Our results contradict that analysis. Particularly in July 2001 there is marked response in surface conductance but less so in evaporation. Such a response has also been observed by Lloyd et al. (2002). We suggest that the response in transpiration to changes in g_s may be masked by the interplay between on the one hand a reduction in conductance and on the other an increase in atmospheric demand. This assumes soil evaporation can be neglected.

The forests of the Far East Siberia survive in an extreme environment, low annual rainfall and extremely low temperatures, possibly complemented by strong nitrogen deficiencies. The precipitation varies considerably between years, but as Sugimoto et al. (2003) have shown, water shortages may not occur even in dry years, as an additional supply of fossil water is available that it can be melted from the lower layers in the active root layer. This has a stabilising effect on transpiration. We suspect that the strong vapour pressure deficit response plays a role in keeping the transpiration rates at a level that is sustainable when this extra source of soil moisture becomes available in the late summer. Continuing to transpire at high rates may cause the forest to senesce

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too early in the season. This arguable goes at the cost of a decreased carbon uptake, and thus may be a previously unrecognised factor explaining the low growth rates of the forest in addition to the obvious response of growth to low temperatures. However, the surprisingly high C-uptake rates we observed during the summer appear to provide an additional balancing force that ensures that over the short growing season sufficient carbon is still taken up to provide maintenance for the trees. Remarkably, as shown in Fig. 2, the forest appears to be able to recover from a period of low uptake in July, and increases its uptake of carbon later in the season when conditions are again more favourably.

On the basis of our findings we would suggest that an increase in temperature, leading to higher values of the specific humidity deficit would decrease the CO₂ uptake of these forest. Although, increases on atmospheric CO₂, enhanced precipitation and an extended growing season may counterbalance such a response to climate change, our analysis indicates that the small current net carbon sequestration capacity of Siberian larch forest is extremely vulnerable to these small climatic changes and may show some quite unexpected behaviour in the future.

Appendix: Calculation of CO₂ losses due to respiration trapped soil

After September the permafrost starts freezing on from below and the topsoil gets frozen from above. This leads to a small layer in the soil at temperatures above zero that can still respire. Observations of concentrations of CO₂ (Ivanov, et al., 1995) show indeed very high levels of CO₂. The CO₂ thus trapped is released when the two freezing layers meet and the soil cracks. Observations of CO₂ concentrations in the surface air in December show this as an increase. We can estimate the amount of CO₂ lost in this process by assuming a linear growth of the freezing layers over a period of

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say 75 days from a total soil depth of 1.5 m. This can be expressed as

$$F_{\text{nee}} = 2F_{\text{resp}} \int z(t) dt = 2F_{\text{resp}} \int_{\text{day}=0}^{\text{day}=75} 0.75t - 0.005t^2 dt,$$

where F_{nee} is the total flux of CO_2 lost by respiration, F_{resp} the average respiration rate at zero degrees expressed per $\text{mol day}^{-1}\text{m}^{-1}$. Analysis of night-time respiration against air temperature suggest that this is $0.5 \mu\text{mol m}^{-2}\text{s}^{-1}$ or $0.0288 \text{ mol day}^{-1}\text{m}^{-1}$. The total respiration thus lost would amount to 1.62 mol m^{-2} or $19.44 \text{ gm}^{-2} \text{ C}$.

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Table 1. (a) Monthly mean NEE and day-time and night-time estimates obtained by assuming a day length lasting of 05:00 to 21:00 h (local time) at Spasskaya Pad in 2000 and 2001 (all values in $\mu\text{mol m}^{-2} \text{s}^{-1}$). Night-time u_* correction based on u_* -threshold/ u_* .

Month	NEE ₂₀₀₀	Day ₂₀₀₀	Night ₂₀₀₀	NEE ₂₀₀₁	Day ₂₀₀₁	Night ₂₀₀₁
April				0.30	0.23	0.46
May				0.56	0.55	1.54
June				-3.24	-5.67	2.10
July	-1.99	-4.36	3.21	-2.00	-4.03	2.47
August	-1.34	-3.62	3.67	-0.84	-2.56	3.05
September	1.02	0.49	2.18	0.42	0.22	0.86
October				0.10	0.12	0.07

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Table 1. (b) Mean monthly NEE and day-time and night-time estimates obtained by assuming a daylength lasting of 05:00 to 21:00 h (local time) at Spasskaya Pad in 2000 and 2001 (all values in $\mu\text{mol m}^{-2} \text{s}^{-1}$). Nighttime u^* correction based on temperature.

Month	NEE ₂₀₀₀	Day ₂₀₀₀	Night ₂₀₀₀	NEE ₂₀₀₁	Day ₂₀₀₁	Night ₂₀₀₁
April				0.23	0.23	0.24
May				0.66	0.55	0.90
June				-3.30	-5.67	1.90
July	-2.32	-4.36	2.15	-1.95	-4.03	2.63
August	-2.10	-3.62	1.23	-1.22	-2.56	1.74
September	0.57	0.49	0.76	0.33	0.22	0.58
October				0.62	0.12	1.74

total sequestration over 2001: 12.7 mol/m²/yr

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Table 2. Mean monthly latent heat flux (evaporation) at Spasskaya Pad in 2000 and 2001 (all values in $W\ m^{-2}$).

Month	λIE_{2000}	λE_{2001}
April		8.9
May		20.1
June		59.9
July	66.9	56.2
August	50.4	21.9
September	13.6	2.6
October		0.6

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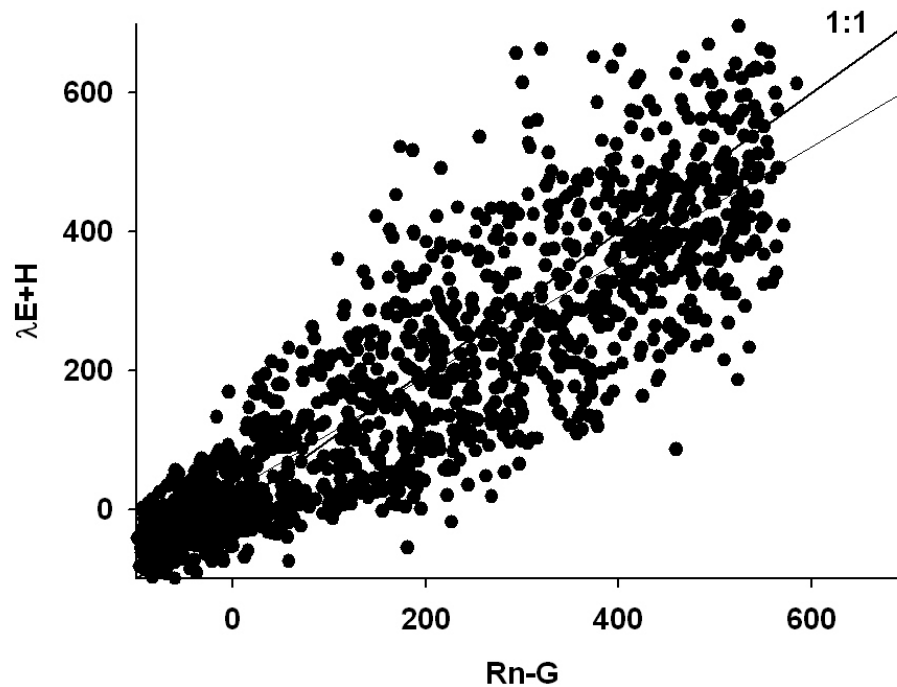
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Fig. 1. Energy balance closure test for 30 min averages of the sum of latent and sensible heat versus net radiation minus soil heat flux. The slope of the line is 0.83 with an intercept of 26 Wm^{-2} and a $r^2=0.79$.

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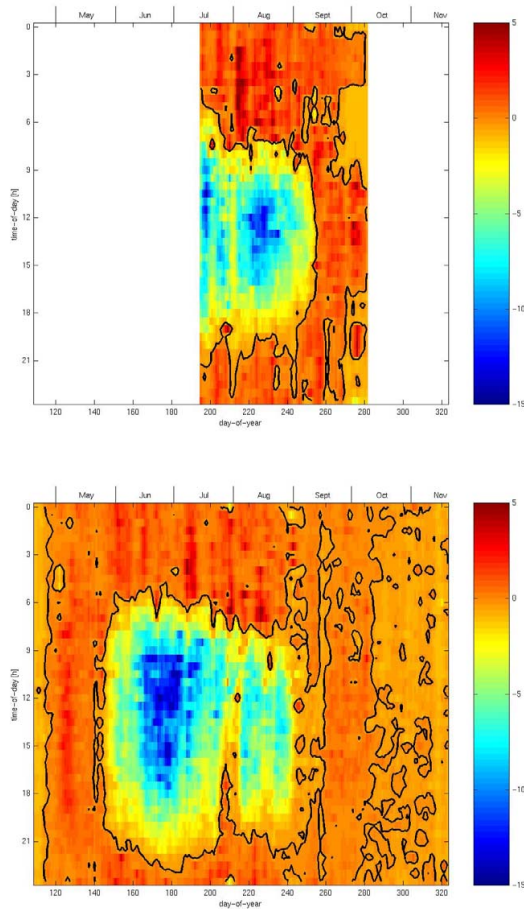


Fig. 2. Diurnal and seasonal CO₂ flux for the larch forest at Spasskaya Pad for 2000 and 2001. Contour colours indicate the value of CO₂ flux ($\mu\text{mol m}^{-2} \text{s}^{-1}$), the line indicates the zero NEE contour.

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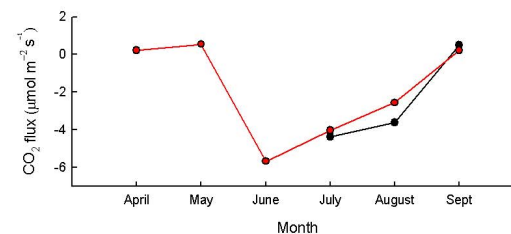
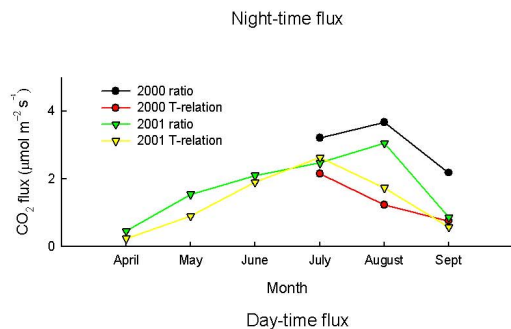
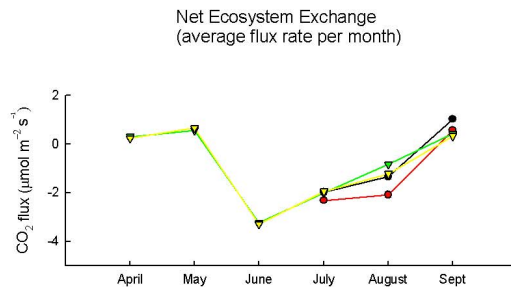


Fig. 3. Monthly averaged CO₂ flux separated in total NEE, and day and night-time flux for 2000 and 2001.

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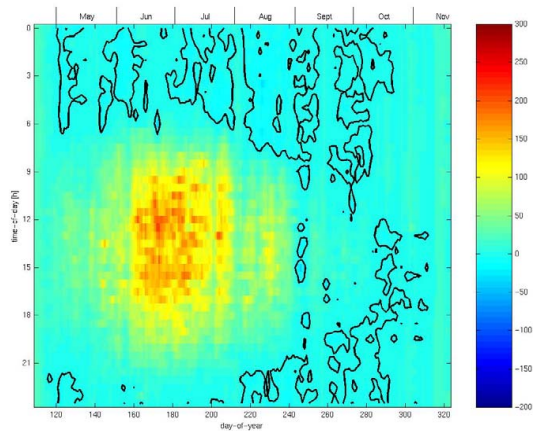
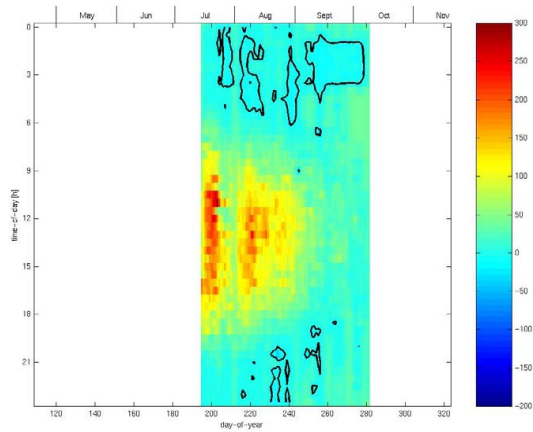


Fig. 4. Diurnal and seasonal course of the latent heat flux above the larch forest at Spasskaya Pad in 2000 and 2001. Contour colours indicate the magnitude of the latent heat flux (Wm^{-2}) and the solid lines indicates the zero latent heat flux contour.

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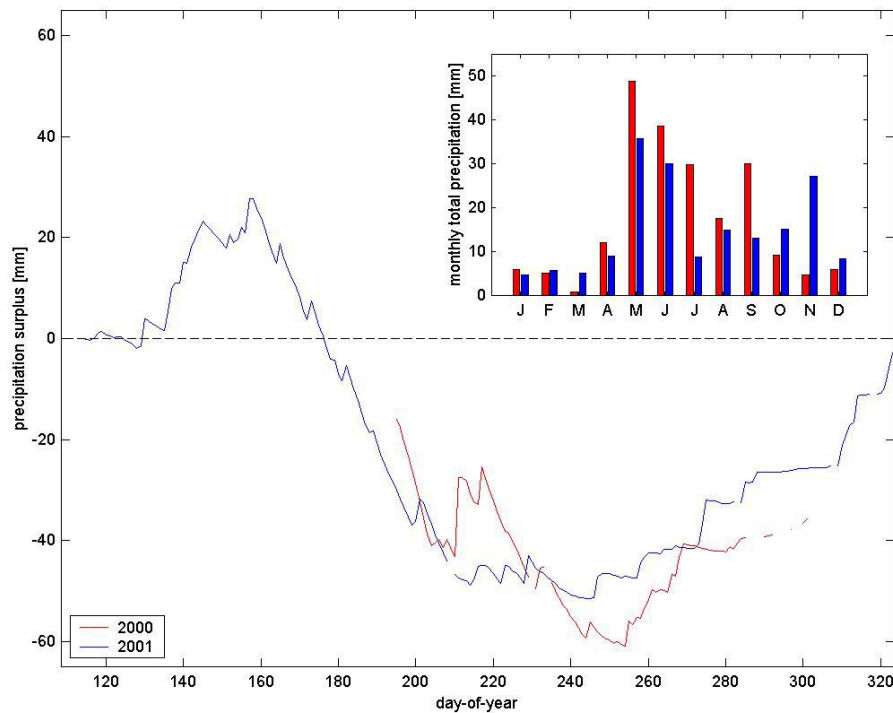


Fig. 5. Running water balance (P-E) based on daily-observed precipitation and evaporation.

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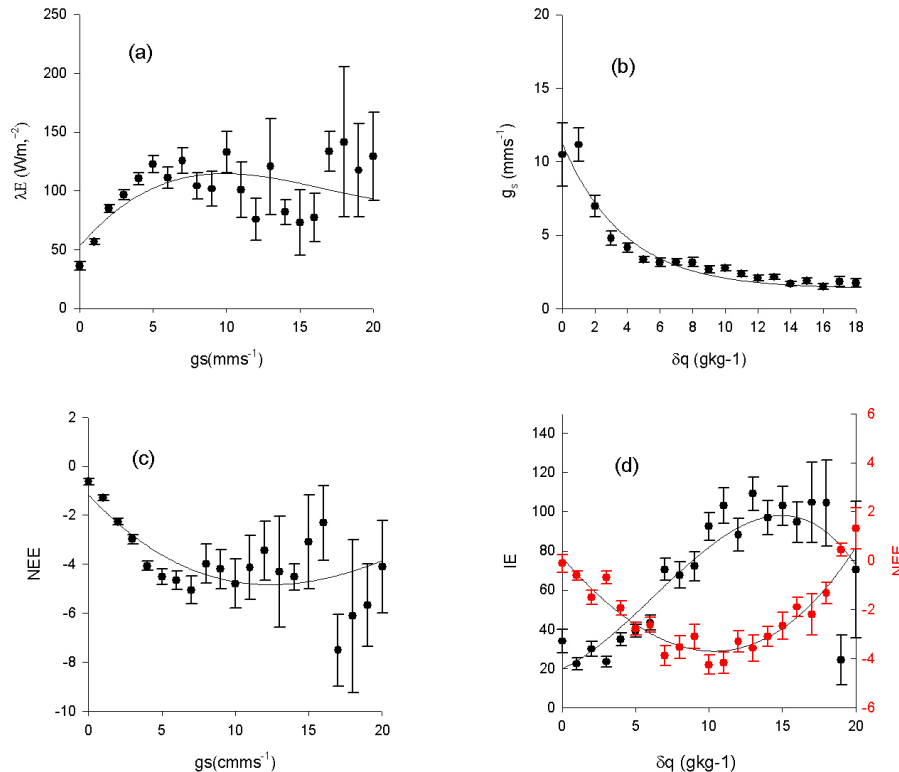


Fig. 6. Relationships between latent heat flux and canopy conductance **(a)**, canopy conductance and specific humidity deficit **(b)**, NEE and canopy conductance **(c)** and NEE and latent heat flux and specific humidity deficit **(d)**. All values are from 2001 and binned. The curves are drawn to guide the eye. Standard errors are shown to indicate the spread within a binned class and thus depend on both the range and number of values.

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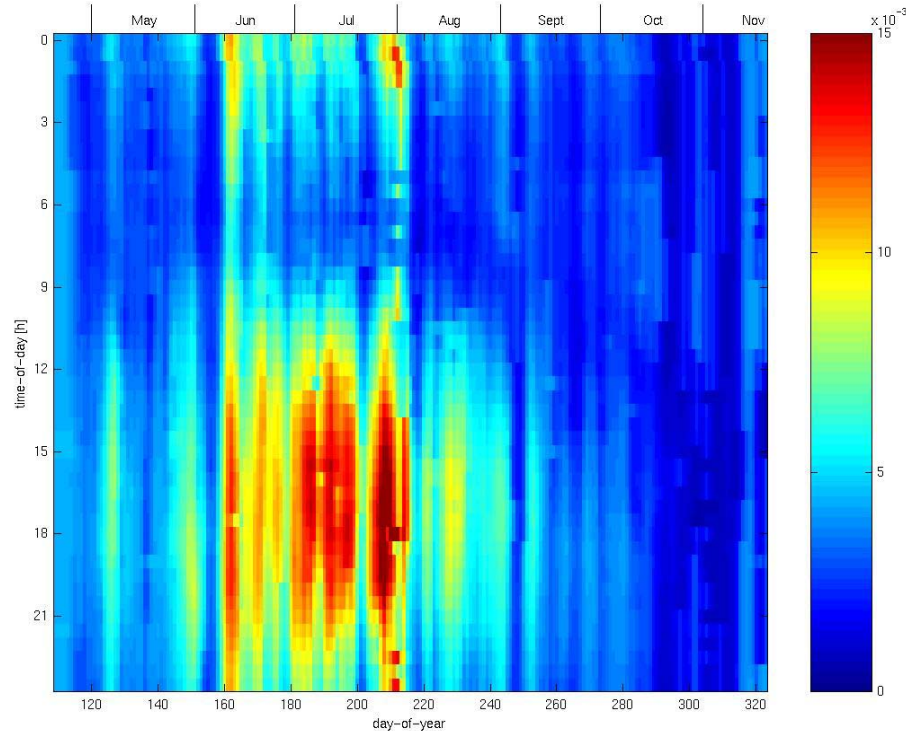


Fig. 7. Diurnal and seasonal course of specific humidity deficit above the larch forest at Spasskaya Pad in 2001. Contour colours indicate the value of specific humidity deficit (g kg^{-1}), the contour line indicates a value of $\delta q = 0 \text{ g kg}^{-1}$.

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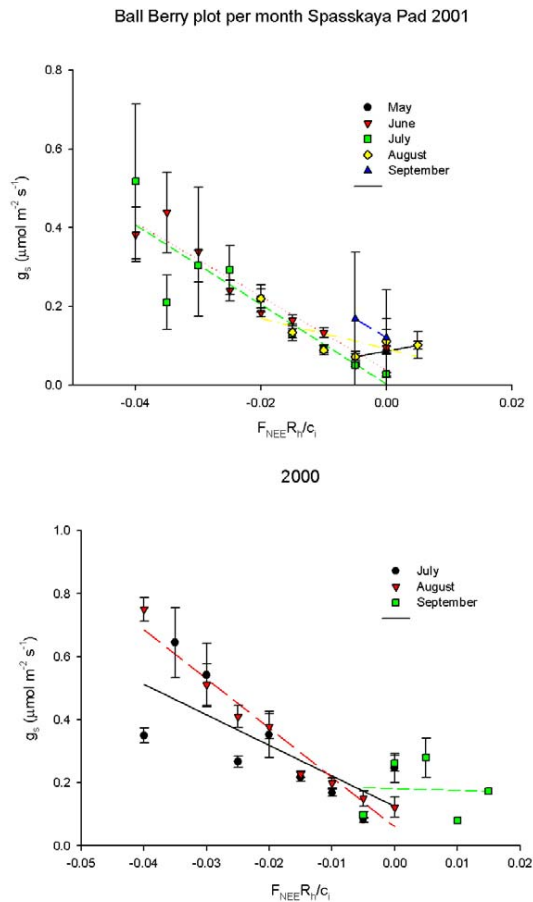


Fig. 8. “Ball Berry” plots for monthly values of conductance and NEE. These illustrate the relations between normalised NEE and canopy conductance as in Ball et al. (1987), but now expressed as a big leaf approximation.

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