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Eddy covariance measurements of CO_2 and sensible and latent heat fluxes during a full year in a boreal pine forest trunk-space

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The eddy covariance method was used to determine turbulent heat fluxes and CO_2 flux below a boreal Scots pine canopy 3 m above the forest floor. Data were filtered using standard deviation of the vertical velocity as a measure of the turbulent mixing in the trunk space along with the non-stationary criterion. The turbulent transfer in the trunk space was dominated by large (15–100 m) intermittent eddies, which were detectable by the eddy covariance technique. Heat fluxes exhibited clear annual and diurnal course and amounted to 20%–30% of the fluxes above the canopy. The forest floor was a source of carbon all-year-round and the CO_2 efflux was mainly controlled by soil temperature. Photosynthesis of the forest floor vegetation decreased daytime fluxes of CO_2 by 1.0–1.5 μmol m^{-2} s^{-1} compared with nocturnal values (~3.0 μmol m^{-2} s^{-1}). The eddy covariance method provided a similar daily cycle to the chamber method but there was a discrepancy between their mean levels.

Introduction

Boreal forests cover a large portion of the northern hemisphere and have a great influence on global climate since they are thought to be one of the major sites of global carbon sequestration (Ciais *et al*. 1995, Tans *et al*. 1995). The net exchange of energy or carbon between a forest ecosystem and the atmosphere is a result of many different sources and sinks functioning in different parts of a canopy. Therefore measurements in different parts of the ecosystem are needed in order to quantify the response of a forest ecosystem to changing environmental conditions. The importance of forest soil and forest floor vegetation in a forest-scale energy and carbon budget depend heavily on the stand structure. Many authors (Black and Kelliher 1989, Baldocci and Vogel 1996, Black *et al*. 1996, Kelliher *et al*. 1997) found that forest floor evapotranspiration in a relatively open-canopied boreal forest can vary between 20%–60% of the total ecosystem evapotranspiration. In sparse forests, the turbulent transport is efficient and there is a lot of available energy in the trunk space below the overstory canopy.

Turbulence inside a forest canopy is intense but intermittent. Although it has been investigated comprehensively during the last decades by several authors (e.g. Raupach and Thom 1981, Amiro and Davis 1988, Baldocci and Meyers 1988, Amiro 1990, Finnigan 2000) no universal theories on sub-canopy turbulence exist. Turbulent transport inside a forest canopy is dominated by strong intermittent gusts that penetrate the canopy and reach the forest floor (Lee and Black 1993, Kaimal and Finnigan 1994, Finnigan 2000). This along with a complex source and sink distribution make traditional micrometeorological approaches such as gradient method (K-theory) inapplicable inside a forest (Raupach and Thom 1981, Kaimal and Finnigan 1994, Finnigan 2000). Thus, vertical fluxes have to be determined by direct methods.

Eddy covariance (EC) is a direct micrometeorological method which allows us to measure turbulent momentum, heat or gas exchange between the underlying surface and the atmosphere *in situ* with only a minimal impact to the local environment. The technique has been routinely used in flux measurements in a constant flux layer since the late seventies. Nowadays, it is applied to more complex situations with good results (Baldocci 2003). The first EC measurements of CO_2 exchange in sub-canopy were made in the mid-1980s (Baldocci *et al*. 1986) and after that the method has been used in several studies concerning energy and CO_2 exchange between a forest floor and the atmosphere (Baldocci and Meyers 1991, Baldocci and Vogel 1996, Blanken *et al*. 1998, Constantin *et al*. 1999, Law *et al*. 1999). However, few long-term experiments that consider the annual variability in energy and CO_2 fluxes have been made (Black *et al*. 1996, Baldocci *et al*. 1997, Blanken *et al*. 2001). Although the EC technique has potential and various authors have used it with promising results, the use of the method is not without its difficulties. While the EC method can provide a spatially averaged estimate of the net exchange of energy or carbon dioxide, chamber techniques can produce important information of the spatial variability of the factors and processes controlling that exchange.

In this paper we describe a full year of EC measurements of sensible heat, latent heat and CO_2 exchange in the trunk space of a boreal pine forest. The goals for this study are: (1) to test and analyze the applicability of the eddy covariance technique for measuring energy and gas exchange in the trunk space and discuss the problems and uncertainties concerning the method, (2) to report the annual and diurnal variability in trunk-space fluxes and investigate the role of the forest soil and forest floor vegetation in energy and carbon fluxes, and (3) to compare the eddy covariance results of CO_2 exchange with chamber measurements and consider the processes controlling trunk-space CO_2-flux.

Materials and methods

Site description

The SMEAR II station is located in a homogenous Scots pine stand (*Pinus sylvestris*) sown in 1962 next to the Hyytiälä forest station of the University of Helsinki in southern Finland (61°51´N, 24°17´E, 181 m above sea level). According to the Cajander site class system (Cajander 1909), the stand is of medium quality and has a growth rate of 8 m^3 ha^{-1} yr^{-1}. The forest is half way through the rotation time for this type of stand and the regeneration has been carried out according to standard silvicultural guidelines (Peltola 2001). The total leaf area index (LAI) is 6 m^2 m^{-2} concentrating in the upper part (6–14 m above the surface) of the canopy leaving the lowest part relatively open (for details of the vertical distribution *see* Rannik *et al*. 2003). The average distance between the trees is 1.5–2 m leading to tree density of 1100–1200 ha^{-1} (Vesala *et al*. 2005). The mean height of the stand was 14.6 m and the mean diameter at breast height 16.2 cm. The stand is homogenous on the eastern sector (0°–180°) up to 200 m whereas on the western sector (180°–360°) it varied between 50–100 m from the mast where an above-canopy eddy covariance setup is located. The site is subject to moderate height variation. Rannik (1998) described the site in micrometeorological context.

The mean annual temperature over the 1960 to 1990 period was 2.9 °C and the precipitation 709 mm. The mean depth of the organic layer is 5.4 cm and density 0.13 g cm^{-3}. The forest floor vegetation is relatively shallow and is dominated by dwarf shrubs and mosses. Average total dry mass, within a 50 m radius from the sub-canopy eddy covariance setup, was 156 g m^{-2} (summer 2004). Of the total dry mass, 17.5% is lingonberry (*Vaccinium vitis-idaea*) and 12.5% blueberry (*V. myrtillus*). The predominant moss species are *Pleurozium schreberi* (48%) and *Dicranum polysetum* (11%). Within a 200 m radius from the sub-canopy EC mast, the contribution of *P. schreberi* is less (18%) and the total dry mass is also smaller (93 g m^{-2}). Along with the shrubs there are some smaller (0.5–1 m high) trees such as birch (*Betula pubencens*). After thinning during winter 2002 (Vesala *et al.* 2005), most of the cut residue was left on the ground.

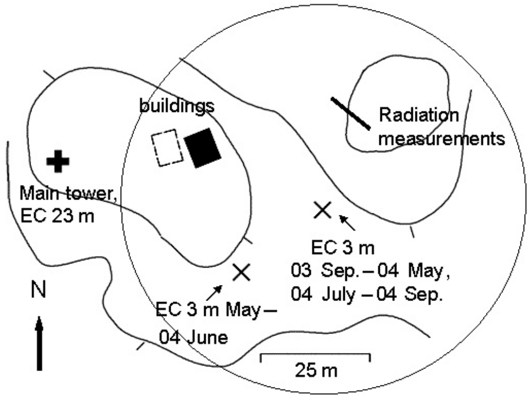

Fig. 1. Measurement site and the location of different measurements. The flux footprint area (80%) for trunk-space fluxes is within the large circle. Interval of the contour lines is two meters. EC refers to eddy covariance measurements at respective height.

Long-term eddy-covariance measurements

Continuous eddy covariance (EC) measurements in the sub-canopy trunk-space were started at the beginning of September 2003. During the one-year period we report in this study, we used two different EC setups each consisting of a fast ultrasonic 3-D anemometer and a closed-path infra-red absorption gas analyzer. Setup A was used from September 2003 to February 2004 and from July to August 2004. It included a Solent 1912R2 (Gill Instruments Ltd, Lymington, UK) anemometer and LI-6262 (Li-Cor Inc., Lincoln, NE, USA) gas analyzer. Setup B operated over the February to June 2004 period. A Metek USA-1 (Metek GmbH, Germany) anemometer and LI-7000 (Li-Cor Inc., Lincoln, NE, USA) gas analyzer were used in this setup. Measurement systems were mounted 50 to 60 m southeast from the main tower three meters above the soil surface (Fig. 1). The gas samples were taken 10 cm below the anemometer and drawn along a Teflon tube (36 m in length and 7 mm in diameter) to the gas analyzer at a flow rate of 16 l min^{-1}. The flow rate was chosen to enable the flow in the tube to remain turbulent (Re ~3500) which minimized the damping of high-frequency fluctuations along the sampling line. We heated the sampling tube to avoid the condensation of water vapor and minimized pressure fluctuations in the sampling system by using a buffer volume between the pump and the gas analyzer as suggested by Aubinet *et al.* (2000). Wind components, virtual velocity and gas concentrations were sampled at 10 Hz frequency and stored into half-hour raw data files.

The above-canopy fluxes were measured using a Solent 1012R (Gill Instruments Ltd., Lymington, UK) anemometer and LI-6262 (Li-Cor Inc., Lincoln, NE, USA) gas-analyzer on the main tower 23 m above ground (about 9 m above tree tops) (Fig. 1). Vesala *et al.* (1998) and Markkanen *et al.* (2001) described the above canopy EC measurements in more detail. We calculated the flux footprints inside and above the canopy using a one-dimensional Lagrangian stochastic trajectory model (Rannik *et al.* 2003) for neutral stratification. Inside the canopy 80% of the total flux originates within 50 m and above the canopy within 200 m in upwind direction.

Flux calculation

We calculated the half-hour mean fluxes using commonly accepted methods (so-called Euroflux methodology, Aubinet *et al.* 2000). We determined the mean value ($\overline{s}$) for each variable as 30 min block-average and the momentary fluctuation as $s' = s - \overline{s}$. To minimize the effect of sloping ground and any possible misalignment of the anemometer to the fluxes (Baldocci and Meyers 1991, Kaimal and Finnigan 1994), we applied a three dimensional co-ordinate rotation for the wind components. In this process, the *u* component is rotated along the mean wind direction and the mean vertical wind speed ($\overline{w}$) and the covariance between vertical and lateral wind fluctuations ($\overline{w'v'}$) are forced to zero (Kaimal and Finnigan 1994). After the co-ordinate rotations, we calculated mean vertical fluxes of sensible (H) and latent heat (LE) and CO_2 (F_c) using the equations

$$H = \rho c_p \overline{w'T'} \quad (1)$$

$$\mathrm{LE} = L\rho \overline{w'q'} \quad (2)$$

$$F_c = \rho \overline{w'c'}. \quad (3)$$

In these equations, ρ is the density of the air, c_p the heat capacity at constant pressure, T' virtual temperature fluctuation, q' and c' the fluctuations in H_2O- and CO_2-mixing ratios, L the latent heat of vaporization and overbar denotes time averaging. We define upward fluxes as positive. A time lag between w' and q' or c', caused by a delay in gas sampling line, was calculated by synchronizing w' and s' using maximum correlation method (Aubinet *et al*. 2000). We removed the effects of humidity and lateral momentum flux perturbations on the sensible heat flux as suggested by Aubinet *et al.* (2000).

The frequency interval measured with an EC setup is limited at the low frequency end by the length of the averaging period and at the high frequency end by sensor response causing the measured turbulent flux, $(\overline{w's'})_m$, to be smaller than the real turbulent flux, $\overline{w's'}$. In a closed-path system high-frequency attenuation of the scalar concentration in the sample line and sensor separation are the two most significant reasons for flux underestimation (Aubinet *et al.* 2000). It is common practice to correct measured fluxes for these effects by multiplying them with a correction factor CF, which is defined as the ratio between the real and measured flux (Moore 1986, Aubinet *et al*. 2000)

$$\mathrm{CF} = \frac{\overline{w's'}}{\left(\overline{w's'}\right)_m} = \frac{\int_0^\infty C_{ws}(f)\,df}{\int_0^\infty \mathrm{TF}(f)\,C_{ws}(f)\,df}. \quad (4)$$

The correction factor depends on the system characteristics, wind speed and atmospheric stability. In Eq. 4, $C_{ws}(f)$ is co-spectral density, f is frequency and TF(f) is a transfer function characterizing the measurement process. TF(f) can be estimated using theoretical (Moore 1986), analytical (Horst 1997, Massman 2000) or empirical approach (Aubinet *et al*. 2000). Calculation of the CF needs also knowledge of the shape of the real, undisturbed cospectrum or an accepted cospectral model which currently does not exist for sub-canopy turbulence.

During the summer of 2005 we made an attempt to quantify CF using simultaneous concentration data measured with an open-path gas analyzer (LI-7500, Li-Cor Inc., NE, USA) and a closed-path analyzer (LI-7000). We attached the open-path to setup B, made a WBL correction (Webb *et al.* 1980) to the concentration signals and compared the cospectra of CO_2 and H_2O with the temperature cospectrum. All open-path cospectra showed similar frequency behavior and therefore we used an empirical approach to determine TF(f) for setup A and B. We assumed a similarity between temperature and scalar transport and calculated the ratio of normalized cospectral densities of a scalar *s* and temperature. After that, we fitted a cospectral transfer function as suggested by Horst (1997)

$$\mathrm{TF}(f) = \frac{1}{1 + \left(2\pi f \tau_c\right)^2} \quad (5)$$

to the data and determined the effective time constant τ_c. The details of the method can be found in Aubinet *et al.* (2000) and will not be repeated here. We estimated CF's for CO_2 and H_2O using Eq. 4 with several different reference $C_{ws}(f)$'s. However, since the corrections were relatively small and the scientific basis still relatively unknown (in trunk space), we chose not to apply these corrections to the fluxes that we obtained in this study.

Additional measurements

Air temperature and moisture were measured continuously on several levels between 4.2 m and 67.2 m with PT-100 sensors (platinum resistance thermometers) and a URAS 4 gas analyzer (Hartman & Braun, Frankfurt am Main, Germany). Soil temperature was measured at several depths below 4 cm with silicon temperature sensors (Phillips KTY81-110). Later in this study, the air temperature (T_a) refers to temperature at 4.2 m height and the soil temperature (T_s) to temperature at 5 cm depth. Photosynthetic photon flux density (PPFD) at the forest floor was measured with four PAR sensors (Li-Cor Inc., Lincoln, NE, USA) mounted on a 4-m-long boom 70 cm above the soil surface. In autumn 2004, we started net radiation measurements in the trunk space using three radiation sensors (MB-1, Tartu Observatory, Estonia). All radiation measurements were made at a fixed location about 30 m northeast from the below-canopy eddy covariance setup (Fig. 1) at one minute intervals.

Alongside the EC measurements, the CO_2 flux from the forest floor was also measured with a chamber system. This system consists of three automated chambers (diameter and height 20 cm) which are connected to infrared gas analyzers (URAS 4, Mannesmann, Hartmann & Braun, Frankfurt am Main, Germany) via heated PTFE tubing (length 40 m, diameter 6 mm/4 mm). A flow of compensation air with known CO_2 concentration was introduced into the chamber at a flow rate equal to that with which the sample air was pumped into the analyzer. The flow rates of the compensation air and the sample air were controlled by separate pumps and mass flow controllers. Chambers were closed for 240 seconds once an hour. The flux was calculated using the concentrations of the inflow and the outflow and the flow rate. The chambers were transparent and the vegetation inside the chambers was left intact. Therefore, the measured CO_2 flux is a combination of CO_2 efflux from the soil and the photosynthetic CO_2 uptake by the vegetation. The chamber system is described in detail by Pumpanen *et al.* (2001).

Data quality analysis and data selection criteria

We filtered the fluxes using the non-stationary criterion suggested by Foken and Wichura (1996). Accordingly each half-hour period was divided into six sub-periods, fluxes were calculated separately for each sub-period and, finally, the difference between the average 'sub-flux' and the mean half-hour flux was compared with the mean half-hour flux. We considered the fluxes as stationary when the absolute value of this difference was smaller than unity (< 100%). Aubinet *et al.* (2000) defined surface layer EC measurements to be of high quality when the difference is smaller than 30% and of acceptable quality when the difference is between 30% and 60%. However, we observed no significant difference between 60% and 100% thresholds in trunk-space fluxes. This test filters out situations when flux magnitude or direction varies highly within the averaging period.

In calm conditions, the measured value of CO_2 flux does not represent the actual exchange between the underlying surface and the atmosphere (Goulden *et al.* 1996, Aubinet *et al.* 2000). When the mean wind speed was low, F_c varied greatly in magnitude and in direction (Fig. 2a). When we plotted F_c against the standard deviation of the vertical velocity (σ_w), we observed that above a certain threshold value of σ_w, most of the variation vanished and the flux was only very weakly dependent on σ_w (Fig. 2b). Interestingly, this method — qualitatively similar to friction velocity threshold (e.g. Foken and Wichura 1996, Aubinet *et al.* 2000) — saved some observations made under very weak mean winds (Fig. 2c). The threshold value of σ_w for CO_2 flux seemed to vary from 0.07 m s^{-1} in setup A to 0.11 m s^{-1} in setup B. Based on Fig 2., we used σ_w as a measure of the amount of turbulence present in the trunk space and neglected CO_2 fluxes collected during periods of insufficient mixing. We did not apply the σ_w-mixing criterion to the H_2O flux (LE) since under calm nights LE remained near zero and showed very weak dependence on the σ_w.

The topography of the site is not ideal for eddy covariance measurements and the effect of sloping ground should be taken into account in

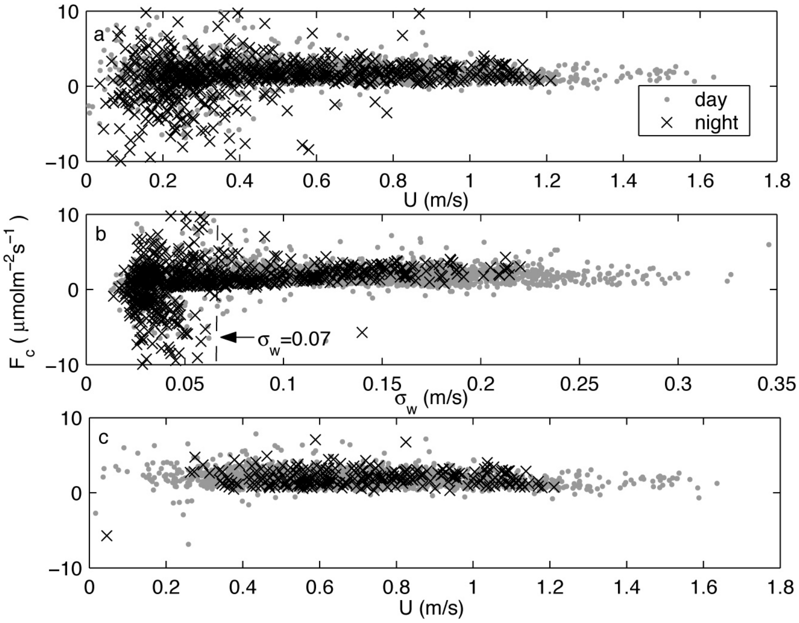

Fig. 2. CO_2 flux measured 3 m above the forest floor (F_c) in a Scots pine forest. (**a**) F_c as a function of mean wind speed (U). (**b**) F_c as a function of the standard deviation of the vertical velocity (σ_w) which can be used as a measure of turbulent mixing in the trunk-space. (**c**) F_c as a function of U when σ_w-mixing criterion has been used. Daytime and nighttime observations are separated using sun elevation angle less than 3° as a limit.

trunk-space fluxes (Baldocci and Meyers 1991). In the northern sector, just inside the footprint area, there is a small building and a measurement tower (Fig. 1) that theoretically could disturb the measurements when the wind blows from that particular direction. However, the wind direction had no significant effect on the actual measured fluxes. Therefore, we chose to accept all the data regardless of the prevailing wind direction.

In the summer of 2004 we compared the two EC setups (A and B) side-by-side (horizontal distance 1 m) in the field. The agreement between the two setups was good. Regression analysis for H and LE showed that H measured by setup B was on average 2% lower than H measured by setup A (r^2 = 0.86, slope = 0.92, intercept 0.4 W m^{-2}). For LE the difference was 1% (r^2 = 0.82, slope = 0.99, intercept 2 W m^{-2}). Regression of the CO_2 fluxes was poorer, F_c measured by setup B was on average 20% higher than F_c measured by setup A (r^2 = 0.67, slope = 1.2, intercept –0.36 µmol m^{-2} s^{-1}). This may indicate a better frequency response of LI-7000 gas analyzer compared to LI-6262. The 20% difference in half-hour CO_2-fluxes between the setups was in the range of uncertainty of an individual flux measurement (Wesely and Hart 1985, Businger 1986, Wilson and Meyers 2001) and therefore we did not adjust any of the fluxes.

Results

Spectral analysis

The power spectrum of vertical velocity showed a broad peak for those frequencies corresponding to eddy sizes 10–20 m and the slope at higher frequencies was less steep than that for surface layer turbulence (Kaimal and Finnigan 1994). The power spectra of horizontal wind components peaked at eddy sizes 50–150 m. Horizontal wind components exhibited a spectral short cut where energy from large eddies was transformed directly into small eddies due to the interaction between the large, shear or buoyancy created, eddies and the rough canopy (e.g. Finnigan 2000). In general, the power spectra displayed typical characteristics of within canopy turbulence (Amiro and Davis 1988, Blanken *et al.* 1998, Kaimal and Finnigan 1994).

The normalized cospectrum between vertical wind and temperature showed nearly the same frequency behavior as CO_2 and H_2O cospectra (Fig. 3). At low frequencies the differences can be explained by the large uncertainty in $C_{ws}(f)$. The closed-path CO_2-cospectrum attenuated at high frequencies relative to the open-path and model cospectra (Kaimal *et al.* 1972 for near-neutral stratification) (Fig. 3) causing the measured flux

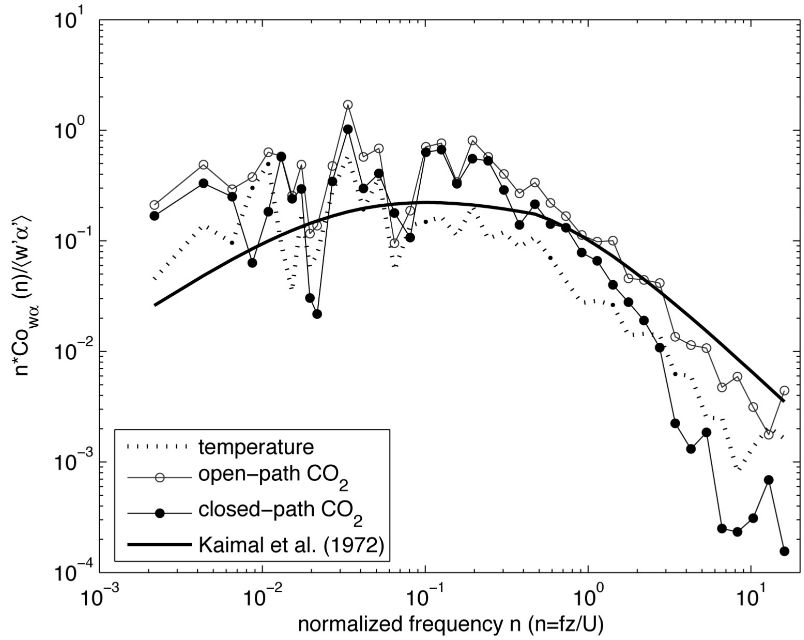

Fig. 3. Typical sub-canopy cospectra of vertical wind and temperature and vertical wind and CO_2 measured with a closed-path and an open-path. Closed-path cospectrum attenuates at frequencies $n > 2$ relative to open-path and temperature cospectra. A cospectral model (Kaimal *et al*. 1972) shown as a reference.

to be an underestimate of the actual flux. The attenuation was clearly seen when we compared open-path and closed-path power spectra (not shown). We used the empirical method described earlier to calculate the effective time constants (τ_c) and the experimental transfer function TF(f) (Eq. 5) for both closed-path EC setups. After that, we calculated correction factors (CF) for CO_2 and H_2O fluxes using Eq. 4. We found that τ_c's were of the same magnitude for both the setups with larger mean values for H_2O than CO_2 (0.40 ± 0.10 s and 0.25 ± 0.10 s). In Fig. 4, the model cospectra and open-path CO_2-cospectrum are used as a reference, they are multiplied by TF(f) and compared with the actual closed-path CO_2-cospectrum. The empirical method to determine TF(f) seems satisfactory since both reference cospectra are able to produce the same frequency behavior at high frequencies as the closed-path cospectrum. Resulting CF's (Eq. 4) not only depend on τ_c but also on the cospectrum used as a reference. For typical wind speeds (U ~1–2 m s^{-1}) CF ranged from 1.01 to 1.08 (CO_2) and 1.03 to 1.12 (H_2O). Using the model cospectrum (Kaimal *et al*. 1972) as a reference produced the largest values whereas the temperature and the open-path cospectrum gave the lowest CF's. This is because the model cospectrum peaks at higher frequencies than the actual sub-canopy cospectra (Fig. 3) and thus overestimates the importance of high-frequency transport in the total flux.

At high frequencies CO_2 and H_2O signals were (0.2 Hz and 1.0 Hz (setup A), 1.5 Hz and 1.0 Hz (setup B)) dominated by white noise but it did not significantly degrade our measurements since the white noise and w' were not correlated. The lower noise thresholds in setup A reflect the poorer signal-to-noise ratio of the LI-6262 analyzer relative to the LI-7000 analyzer. Experimental cospectra represented in Figs. 3 and 4 were calculated by applying standard FFT routines to linearly detrended and Hamming-windowed (Kaimal and Kristensen 1991) segments of 2^{14} datapoints (about 13 minutes). Each cospectrum is an average of 90 minutes of data collected during a sunny summer day and is plotted against non-dimensional frequency n ($n = fz/U$).

Energy balance closure

The surface energy balance below an overstory canopy can be expressed as

$$R_n = H + \mathrm{LE} + G + S, \qquad (6)$$

where R_n is net radiation, H is sensible heat and LE latent heat flux, G is heat flux into the soil

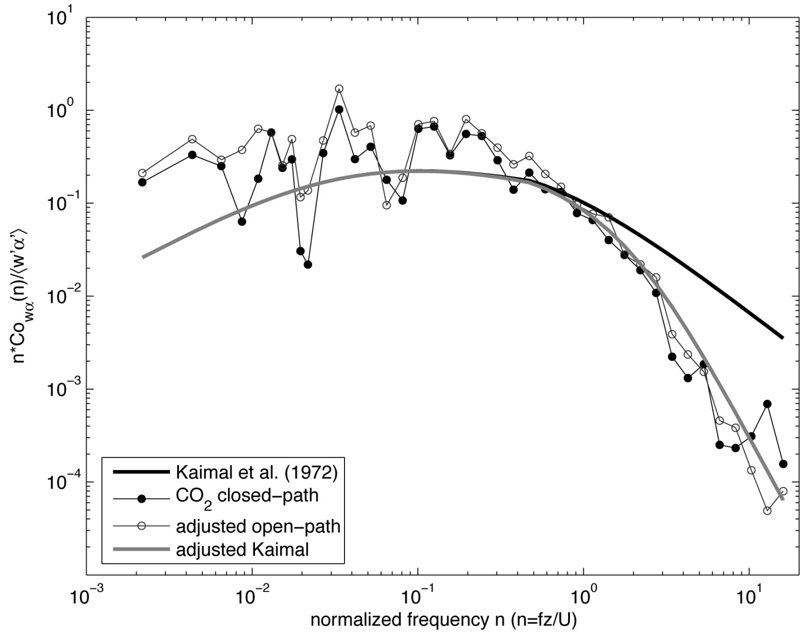

Fig. 4. The frequency behavior of a closed-path CO_2 cospectrum can be achieved by multiplying an open-path, or model cospectrum with a transfer function TF (Eq. 5).

and *S* is heat storage below the observation level. We estimated *S* as the sum of sensible and latent heat storage into the air column between the soil surface and the measurement level using changes in temperature and water vapor concentrations measured by EC. However, we omitted heat storage into biomass when estimating *S*. We calculated soil heat storage using soil temperature and moisture data and estimated *G* to be equal to a change in heat storage in the uppermost 11 cm of the soil. The slope between turbulent fluxes (*H* + LE) and available energy ($R_n - G - S$) in September 2004 was 0.55 ($r^2 = 0.71$, zero intercept) indicating that the energy balance at the forest floor was not closed. The closure was more complete in the morning and late afternoon compared with that observed at midday.

Below-canopy fluxes

Annual patterns

Sensible and latent heat fluxes (*H* and LE) below a Scots pine canopy showed clear annual variation (Fig. 5a and b). Evaporation from the surface was negligible during cold winter months (November to March) and the increase in April was associated with the snowmelt. The midday LE reached its maximum around 100 W m^{-2} in July. During the winter period (November to March) *H* varied typically between –20 and 30 W m^{-2} and the magnitude and the direction were strongly coupled to the stability of the surface layer. The rapid increase in *H* in late April was linked to the increase in soil temperature after snowmelt, which increased the vertical temperature gradient between the surface and overlying air. *H* reached its daytime maximum (~80 W m^{-2}) in late May. The variation in the trunk-space *H* and LE, on a scale of a few days or longer, was driven by the changes in atmospheric weather conditions. This was clearly seen — for example — during a low-pressure situation at the end of July when both heat fluxes remained small over several days. The annual pattern of F_c (Fig. 5c) shows a minimum during the winter (January to March) and maximum during July and August following the annual course of the soil and air temperatures (not shown). The longest gap in F_c (first half of June 2004) and in LE (May and June, Fig. 5) was caused by technical problems with the gas analyzer.

Diurnal variation of the heat fluxes

We calculated an average diurnal course of *H* and LE for the period of maximum assimila-

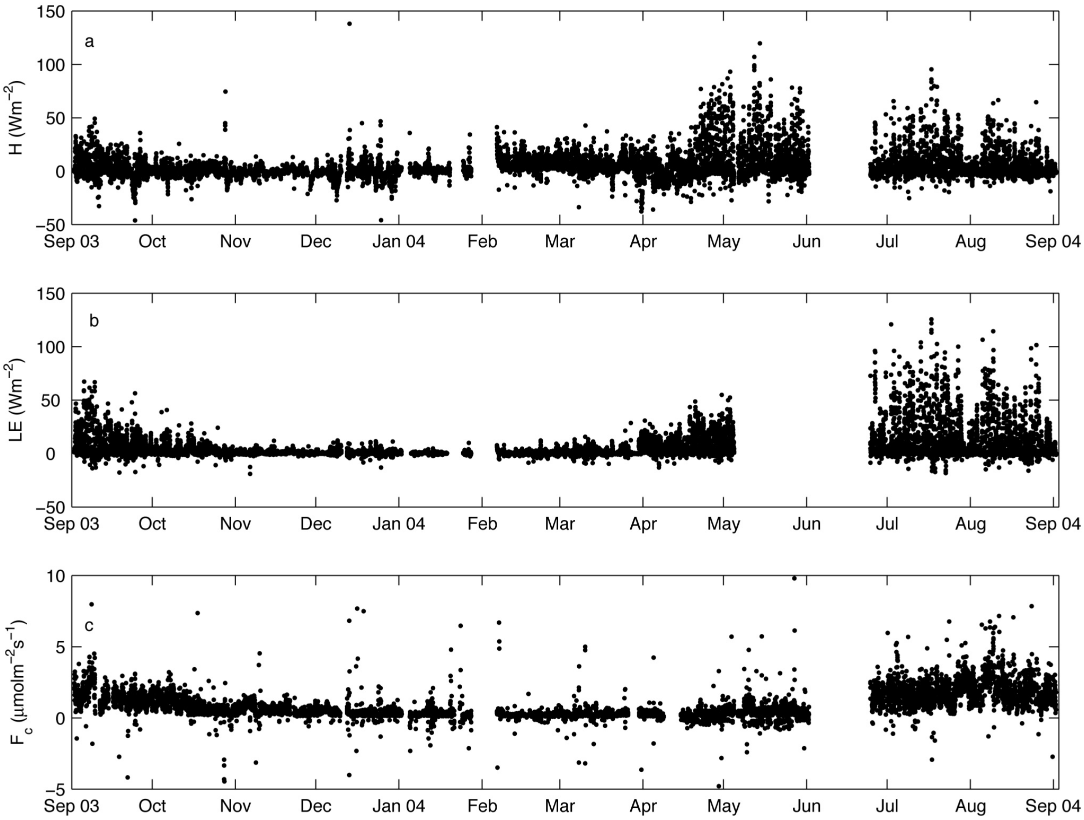

Fig. 5. The annual patterns of (**a**) sensible heat flux *H*, (**b**) latent heat flux LE, and (**c**) CO_2 flux F_c.

tion rate (July and August) of the canopy (for definition of this period *see* Suni *et al*. 2003a). The diurnal variation of sensible heat flux (Fig. 6a) was approximately symmetrical in respect to noon, when it reached its average maximum (30 W m^{-2}). In afternoon (14:00), *H* dropped rapidly and reached a constant level close to zero in the late afternoon (18:00) well before sunset. During the nighttime *H* remained near to zero and rose again after sunrise when solar radiation started to warm the canopy. The latent heat flux (Fig. 6b) showed similar diurnal variation having its maximum (45–50 W m^{-2}) around noon. At midday, the Bowen ratio (β = *H*/LE) in the trunk-space was typically 0.4–0.5. It is worth noting that over the two-month averaging period the diurnal changes in heat fluxes were statistically significant even though the standard deviations were large. The standard deviation of *H* (σ_H) was of the same order as the mean flux indicating strong dependence on the driving environmental variables while during daytime σ_{LE} was somewhat smaller (about 1/3 of the mean).

The comparison between the heat fluxes below and above the canopy is shown in Fig. 6c. The above-canopy midday values of *H* and LE were almost equal (120 W m^{-2}) leading to a Bowen ratio around unity. The surface layer was (on average) hydrostatically stable during nights and the above-canopy *H* was directed downwards (*H* ~ –20 W m^{-2}). This was in contrast to the trunk space where the nighttime *H* remained near zero. The combination of evaporation from the forest floor and understory evapotranspiration was responsible for approximately 30% of the ecosystem scale evapotranspiration (Fig. 6c) whereas the contribution to sensible heat flux was smaller (~20%). Vertical mixing (σ_w) was greatest around noon (Fig. 6d) and followed closely the course of incoming shortwave radiation (not shown). Vapor pressure deficit (VPD),

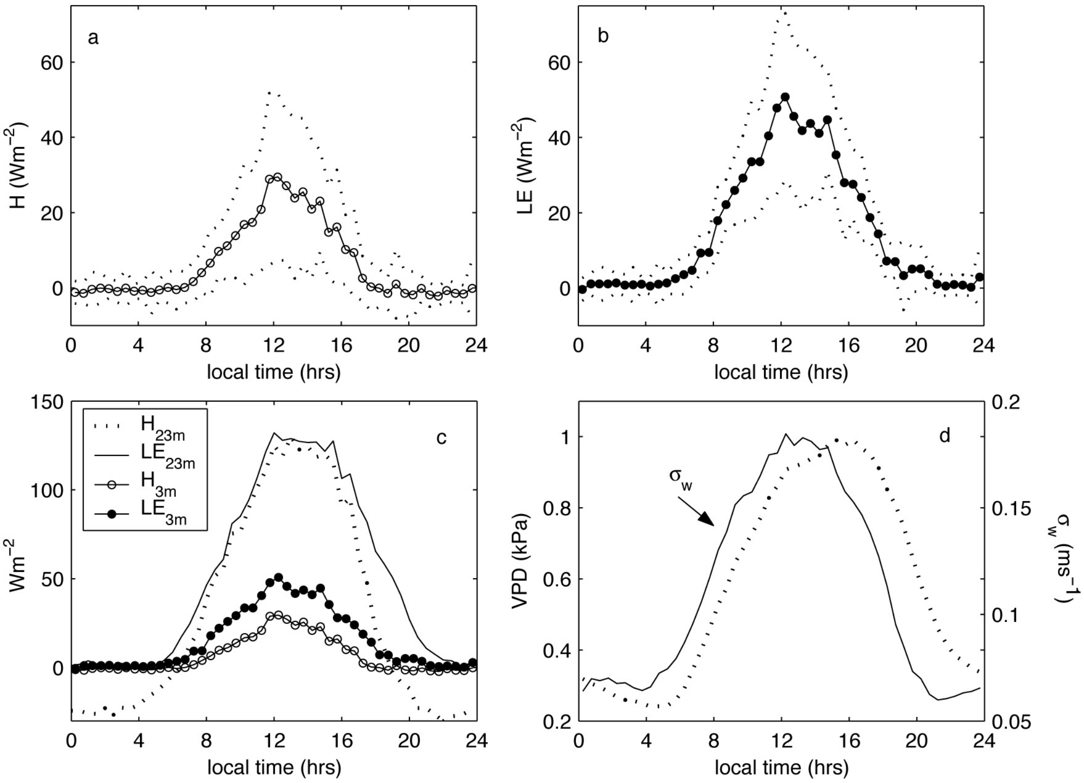

Fig. 6. The diurnal course of heat sensible heat flux (*H*) and latent heat flux (LE) in July–August. (**a**) *H* and (**b**) LE in trunk-space (3 m); mean ± standard deviation is plotted with dotted line. (**c**) *H* and LE in trunk space (3 m) and above the canopy (23 m). (**d**) Diurnal courses of vapor pressure deficit (VPD) and standard deviation of the vertical velocity (σ_w) in trunk space.

instead, was strongly dependent on air temperature which lagged the radiation field (Fig. 6d).

Trunk-space CO_2-flux

Diurnal variation

Over July and August, the diurnal course of the trunk-space F_c (Fig. 7a) varied significantly with minimum (1.5–1.7 μmol m^{-2} s^{-1}) at midday and maximum (2.4–2.5 μmol m^{-2} s^{-1}) in the late evening. Early in the morning (05:00) there was a depression in F_c followed by a rapid increase. We assume that during the relatively calm nights a proportion of the respired CO_2 is stored in the air column between the soil surface and the measurement height and it is "flushed" upwards when vertical mixing is enhanced after sunrise. Understory photosynthesis, together with lower soil temperature as compared with that in the evening, caused a diurnal minimum around noon. In the afternoon F_c started to rise. This was probably due to rising soil temperature and decreasing photosynthesis with decreasing photosynthetic photon flux densities (PPFD). The mean maximum at the late evening occurred when understory photosynthesis was negligible and soil temperature had just passed its diurnal maximum (T_s had a broad maximum around 18:00–20:00. The time lag with respect to air temperature was about two hours). The diurnal variation in July–August was as for the heat fluxes, clear. However, the standard deviation was about half of the mean flux indicating a strong variability with changing environmental conditions such as soil temperature and reflecting a high natural variation in turbulent exchange. During winter (Fig. 7c) F_c was small (0.2–0.4 μmol m^{-2} s^{-1}) and the diurnal variation was negligible.

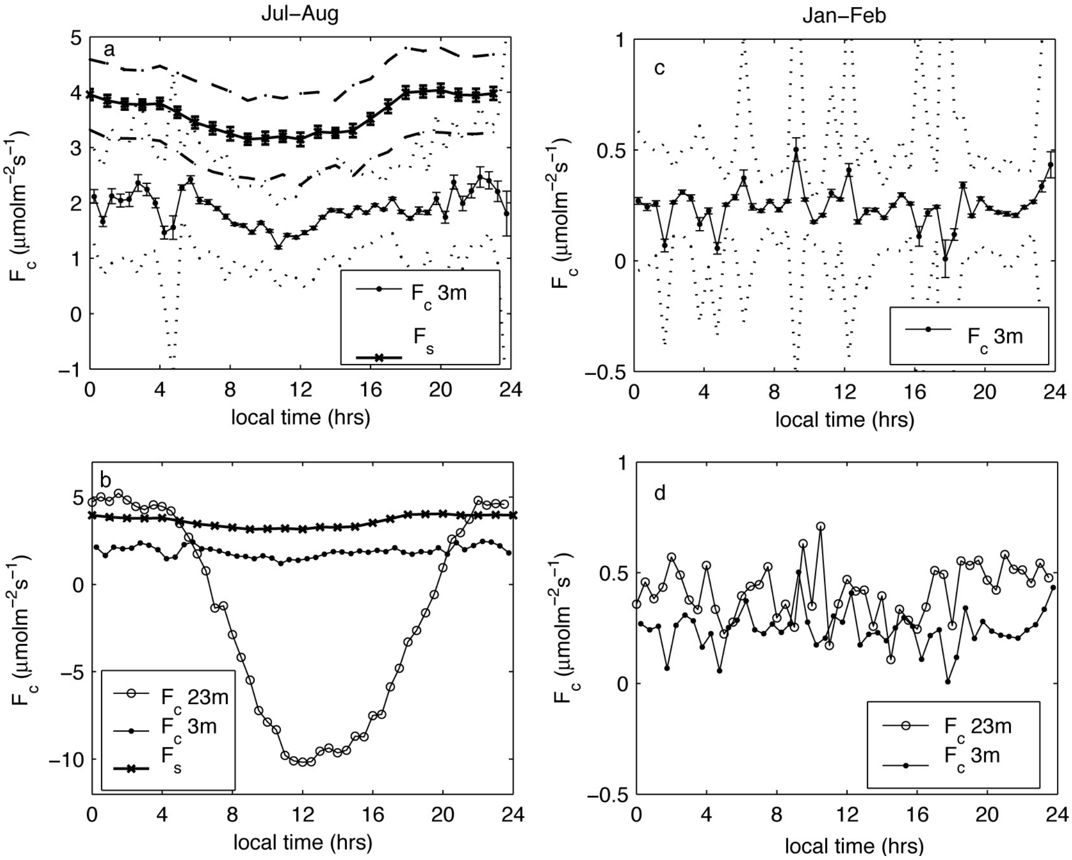

Fig. 7. The diurnal courses of CO_2 fluxes (F_c) in July–August (left-hand side column) and in January–February (right-hand side column). (**a**) F_c in trunk space (3 m) and chamber flux F_s, (**b**) eddy covariance fluxes (F_c) in trunk space (3 m) and above the canopy (23 m) and F_s, (**c**) F_c in trunk space, (**d**) F_c in trunk space and above the canopy. In panels **a** and **c** the standard errors of the mean are plotted with small vertical bars and ± standard deviation with dotted line.

The chamber flux (F_s) displayed similar diurnal course and variability as F_c (Fig. 7a) with maximum values ~4.0 μmol m^{-2} s^{-1} and minimum values 3.2 μmol m^{-2} s^{-1}. However, the magnitude of F_s was about 1.6 to 1.7 times F_c, a difference that could partly be explained by spatial variability in surface characteristics and partly by differences in the methods.

During July and August, the above-canopy F_c (Fig. 7b) exhibited clear nighttime maximum (~5 μmol m^{-2} s^{-1}) and midday minimum (~10 μmol m^{-2} s^{-1}) as expected (Suni *et al*. 2003a). The nocturnal above-canopy F_c should, in turbulent conditions (e.g. $u* > 0.35$ m s^{-1}, Markkanen *et al*. 2001), represent the total ecosystem respiration. Based on eddy covariance measurements (Fig. 7b), we estimate that about 40% of the nighttime ecosystem respiration originates in the forest floor and the understory while the chamber measurements give this proportion to be 75%–80%. In winter (Fig. 7d), the magnitudes of the above- and below-canopy F_c were more equal at midday than during night. This may indicate that during a night we were unable to measure the total CO_2 exchange in the trunk space due to weak, stability inhibited, vertical mixing although the σ_w-mixing criterion was applied.

Factors controlling the trunk-space CO_2-flux

The trunk-space F_c was strongly dependent on the photosynthetic photon flux density (PPFD) showing a tendency to be smaller with increas-

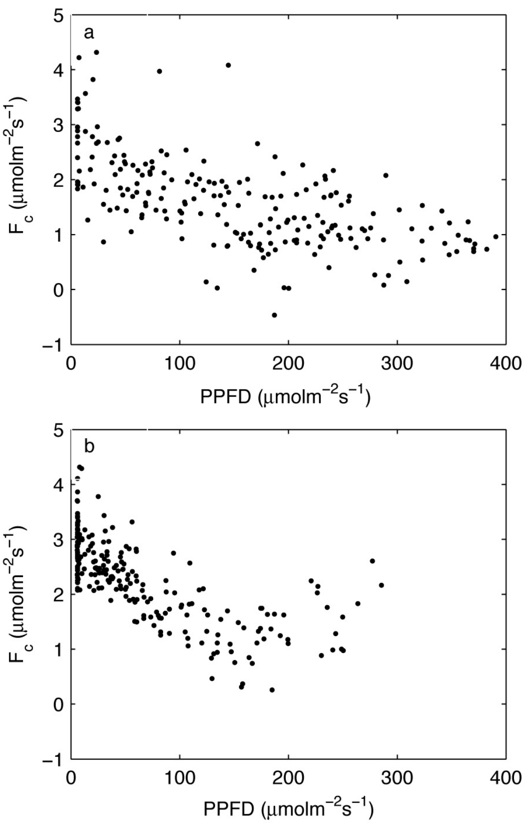

Fig. 8. The relationship between trunk-space CO_2 flux (F_c) and photosynthetic photon flux density (PPFD) in two eight-day periods: (**a**) at the beginning of July, and (**b**) in late July and at the beginning of August. Each point is the average of one hour of data.

ing radiation levels (Fig. 8). At the beginning of July (Fig. 8a) the mean nocturnal F_c was around 2.8 μmol m^{-2} s^{-1} and decreased to a level close to 1.0 μmol m^{-2} s^{-1} when PPFD increased above 200 μmol m^{-2} s^{-1}. However, during a cloudy period in late July and at the beginning of August F_c saturated at the same level but at lower PPFD's (Fig. 8b) and the difference between the mean nocturnal and daytime F_c was around 2.0 μmol m^{-2} s^{-1}. The rapid saturation indicates that the forest floor vegetation was able to photosynthesize effectively at low PPFD's. We estimated, based on EC measurements (Figs. 7a and 8), that the CO_2-uptake rate of the understory was approximately 1.0–1.5 μmol m^{-2} s^{-1} when PPFD was not limiting the photosynthesis. We chose to average PPFD and F_c over one hour (Fig. 8) in order to minimize the effect of any possible unrepresentative radiation data.

We found soil temperature to be the main factor affecting the trunk-space F_c (Fig. 9). The soil respiration and thus F_c increased with increasing soil temperature as expected (e.g. Lloyd and Taylor 1994). In Fig. 9 we separated nighttime observations using calculated sun elevation angle (< –3°) as a limit between nighttime and sunlight hours (Markkanen *et al.* 2001). We used an exponential in the soil temperature model

$$F_c^{\mathrm{mod}} = a\exp\left(bT_s\right), \tag{7}$$

to calculate the quantitative temperature dependency of F_c. Empirical fitting coefficients a and b were 0.28 and 0.14 (mean values of one hundred fits to randomly selected 4/5 of the accepted data). Changes in T_s explained 48% of the total variability in the trunk-space F_c, which indicates that other variables such as the random nature of the turbulent flow, soil moisture content, spatial heterogeneity and physiological effects like photosynthesis play an important role in the trunk-space CO_2-flux (e.g. Constantin *et al.* 1999). We also tested several other models, which included the exponential in air and linear in soil temperature model that Markkanen *et al.* (2001) used for gap filling of the above-canopy F_c dataset. However, none of them provided a better fit than Eq. 7.

Annual net CO_2-exchange in trunk-space

Of the total 12-month dataset (Fig. 5c), 11% of the half-hourly values were missing due to technical problems or measurement breaks and 33% were rejected by non-stationarity or σ_w criteria. Of the missing observations, we filled 91% using the regression model above (Eq. 7) and the remaining 9% using annual mean, when no T_s was available. Figure 10 shows the whole dataset separated into real and interpolated observations. The model (Eq. 7) produced a 1% larger cumulative net carbon exchange on average than the real observations (the remaining 1/5 of the data was used to verify the model). The cumulative net carbon exchange between the forest floor and the atmosphere at 3-m height (Fig. 10b) increased relentlessly over the measurement period. This

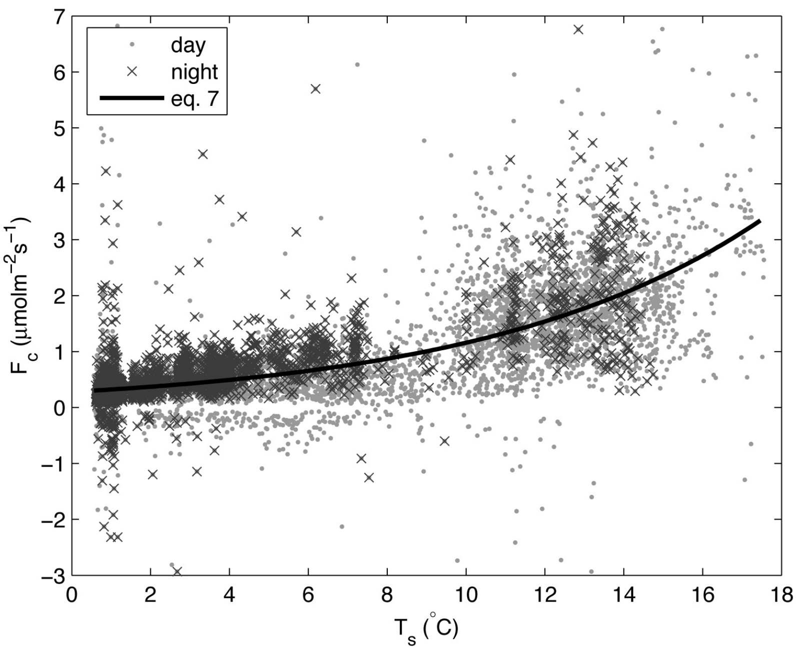

Fig. 9. Trunk-space CO_2 flux (F_c) as a function of soil temperature (T_s) at 5 cm depth. Parameters of the fitted exponential curve (Eq. 7) are: $a = 0.28$, $b = 0.14$.

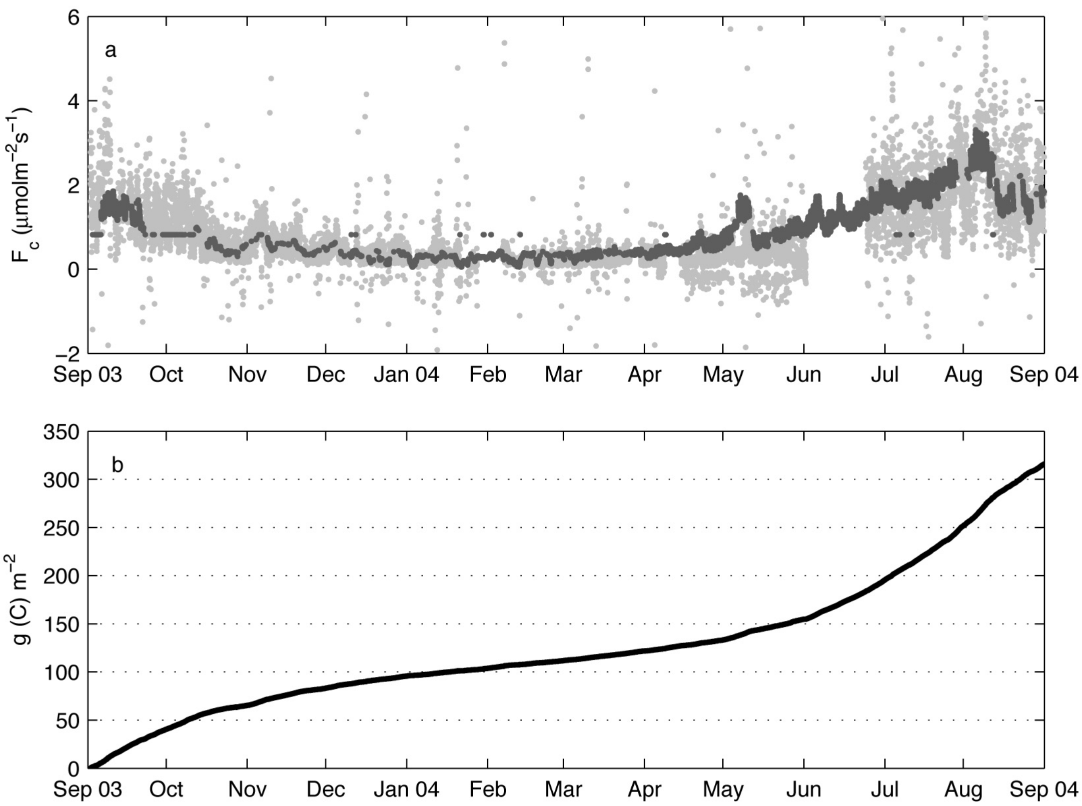

Fig. 10. Annual net exchange of CO_2 between the forest floor and the atmosphere measured by eddy covariance in trunk space. (**a**) Half-hour eddy covariance fluxes (light grey) and observations interpolated using Eq. 7 (dark grey), and (**b**) cumulative carbon exchange (g(C) m^{-2}) from the beginning of September 2003 to the end of August 2004. The forest floor was a source of carbon year-round.

indicated that the soil–understory column was a source of carbon during the whole year. Carbon release from the surface was greatest during the growing season (May to September) although the understory photosynthesis partly compensated the soil respiration. During the winter (November to April), the release of carbon was lesser due to lower soil temperature. During the measurement year, which ran from September 2003 to August 2004, the net exchange of carbon, calculated from the EC measurements was 317 ± 40 g(C) m^{-2}, compared with an estimate of ~535 g(C) m^{-2} from the chamber measurements.

Discussion

The use of eddy covariance method in a trunk space

In this study we tested closed-path eddy covariance instrumentation below a relatively sparse Scots pine canopy. W found that the technique can be used for trunk-space flux-measurements if the prerequisite conditions of the method — stationarity, horizontally homogenous fetch and sufficient turbulent mixing — are met. However, when we apply the method in a sub-canopy trunk-space, where turbulence structure and the source distribution of momentum, heat and trace gases differ from the constant flux layer, where most of the flux measurements are made, some challenges appear (*see* e.g. Baldocci and Meyers 1991).

Within the canopy (inside the roughness sublayer) momentum flux varies with height and the basic surface layer theories do not hold (for surface layer theories *see* e.g. Högström 1996, and for turbulence inside a forest canopy *see* e.g. Finnigan 2000). Although the turbulence intensities would be high below a forest canopy, the turbulence is inactive in the sense of momentum transport (Finnigan 2000) and the friction velocity (u_*, a square root of the momentum flux) does not represent the actual turbulent mixing. We found that correlation between the vertical scalar fluxes and the momentum flux below a forest canopy was poor, an observation which Law *et al*. (1999) have reported earlier. Therefore, from a theoretical point of view, the use of u_* inside the canopy should be avoided and we recommend the use of the standard deviation (or variance) of the vertical velocity as a measure of the vertical mixing in the trunk space. A more practical motivation for the use of σ_w is the fact that inside a forest canopy upward momentum fluxes are occasionally observed (Amiro 1990, Rannik *et al*. 2003) and the friction velocity can then not be defined.

The high frequency attenuation that results from the inability of the measurement system to detect fluctuations associated with small-scale turbulence induces an underestimation of the measured turbulent flux. This low-pass filtering is least for sensible heat and momentum fluxes measured by fast-response ultra-sonic anemometer and largest for closed-path flux systems such as our measurement setups (Massman 2000). For standard EC measurements it is common practice to correct low-pass filtering assuming a similarity between heat and gas transfer and using universal cospectra (e.g. Kaimal *et al*. 1972) as a reference and calculating spectral transfer functions as suggested, for example, by Moore (1986). An alternative approach is to calculate the corrections using analytical methods (Horst 1997, Massman 2000) that are more insensitive to the shape of the cospectra but which need an *in situ* determination or a relevant parameterization for the frequency of the cospectral peak. In a sub-canopy trunk-space, where turbulence statistics and spectral characteristics differ from normal boundary layer conditions, neither the integral nor the analytical methods are necessarily reliable. In the first method, the reference shape of the cospectra may not be representative for sub-canopy turbulence. The use of the second method is complicated by the broadness of the cospectral maxima and potential shifting towards lower frequencies, caused by high frequency attenuation in the measurement system (Massman 2000), making it difficult to determine the exact location of the cospectral peak.

Aubinet *et al*. (2000) suggest an empirical method for the evaluation of the transfer function and the correction factor, which can also be used within the canopy. We followed their approach and used open-path measurements to test if the cospectral similarity is a reasonable approximation in a trunk space. At the site of the study the approximation was indeed found to be reasonable and the resulting empirical transfer functions had lower cut-off frequencies compared to the theoretical transfer function (Moore 1986). Still, the correction factors were at the maximum around 1.10. The model cospectrum (Kaimal *et al.* 1972) overestimates the importance of high-frequency transport giving too large correction factors like Baldocci and Meyers (1991) have suggested. Large intermittent eddies, which dominate energy and mass transport inside a forest canopy, create substantial fluctuations in scalar mixing ratios that are detectable by closed-path analyzers with a high level of accuracy and minimal attenuation.

For almost two decades trunk-space EC measurements have been made by several groups, yet no common practice for quality analysis or flux corrections exists. We therefore emphasize the need for comprehensive testing and comparison of different quality tests and flux corrections in a wide range of canopies in order to achieve a more uniform methodology. For instance, more studies of the shape of the cospectra inside forest canopies are needed.

Energy fluxes in a trunk space

Relatively poor energy balance closure above the forest floor was most likely caused by inadequate net radiation measurements. We observed that at high sun angles (midday) incoming radiation penetrates the canopy through gaps in the foliage making single-point R_n-measurements below the forest canopy non-representative for the whole flux footprint area. Baldocci and Meyers (1991), for instance, observed 100% variation in R_n along a 30 m transect at a forest floor and they therefore stated the need to have horizontally averaged measurements. Typical energy balance closures below a forest canopy range from 0.6 to 1.2 when R_n is measured with a moving device along a path long enough to average R_n over sun and shade patches within the footprint area (Baldocci and Meyers 1991, Baldocci *et al.* 2000, Blanken *et al.* 2001). Above the forest energy balance closures are better (0.7–1.0) according to Aubinet *et al.* (2000), but still more or less unclosed. One of their main interpretations is that horizontal and non-turbulent transport of energy can be an important part of the surface energy budget of complex sites. These effects can be even more pronounced below a forest canopy (Kaimal and Finnigan 1994). At the SMEAR II site, the energy balance closure above the canopy is typically around 0.7–0.8 (Rannik *et al.* 2002, Kolari *et al.* 2004). However, we need to improve R_n measurements at the forest floor to investigate the reasons for the lack of the energy balance closure.

Our results of the forest floor contribution to the total ecosystem evapotranspiration (~1/3) agree well with earlier reports. Sevanto *et al.* (2001) estimated stand-scale evapotranspiration components at the same site as in this study by using trunk-diameter changes, chamber measurements and above-canopy EC measurements. They found that a maximum of 30% of the total evapotranspiration originates from the forest floor and the understory. Using above- and below-canopy EC measurements, Constantin *et al.* (1999) reported that evapotranspiration from the forest floor was about 10% of the total evapotranspiration of a closed-canopied spruce/pine forest during a summer day. Baldocci and Vogel (1996) showed that in a relative open-canopied old-growth ponderosa pine forest, forest floor/understory contribution was between 20%–30% of total stand-scale evapotranspiration, a significantly larger fraction than in a closed-canopied deciduous forest. In an aspen forest evapotranspiration from a forest floor and hazelnut understory was 22% of total evapotranspiration according to Black *et al.* (1996) under a full-leafed period. Black and Kelliher (1989) studied different forest types and found that the fraction of total stand-scale evapotranspiration contributed by the understory varied between 10% and 65%.

The understory contribution to sensible heat flux was smaller than to latent heat flux. We conclude that this was due to the formation of a convective layer on the upper part of the canopy, where a large part of the incoming shortwave radiation is absorbed. At the same time, the canopy prevented the penetration of radiation to the forest floor and decreased the vertical temperature gradients below the overstory canopy. Consequently, H remains small. On the other hand, LE is partly coupled to understory photosynthesis, which can be substantial even at low radiation levels (Goulden and Crill 1997, Whitehead and Gower 2001).

Energy fluxes from the forest floor are a significant part of the total energy budget of a boreal coniferous forest and the contribution depends on the canopy architecture. Lee and Black (1993) concluded that moisture and heat transfer in a Douglas-fir stand is dominated by intermittent dry, cool downdrafts and warm, moist updrafts. At our site the quadrant-hole analysis, in which the momentary products of w' and s' are divided into four quadrants depending on their signs and a momentary value of $w's'$ is compared with the mean flux $\overline{w's'}$ (for description of the method *see* Raupach and Thom 1981, Bergström and Högström 1989), revealed a similar structure. Sensible heat transfer was highly intermittent and dominated by strong events. Typically, half of the total flux was "delivered" in less than 10% of the total time (data not shown). The great importance of the forest floor in stand-scale energy budgets is due to the effectiveness of strong, intermittent gusts to transport the water vapor through the canopy into the atmosphere (Lee and Black 1993, Black *et al.* 1996).

Trunk-space CO_2-flux and the interpretation of eddy covariance and chamber measurements

In turbulent steady-state conditions over a horizontally homogeneous and flat surface, the eddy covariance technique provides a reliable and precise measure of the exchange of momentum, energy or scalar transfer between the underlying surface and the atmosphere (Foken and Wichura 1996, Aubinet *et al.* 2000, Baldocci 2003). However, under more complex conditions, the violation of the assumptions above can cause systematic errors for the interpretation of EC results. Below a forest canopy horizontal advection or dispersive flux (Raupach and Thom 1981, Finnigan 2000) can be a significant part of the total mass or energy balance of the surface, in which case a single-point EC measurement does not represent the actual net exchange between the surface and the atmosphere. Similarly, whether the zero mean vertical velocity that results from the 3-D rotation of the wind components (Kaimal and Finnigan 1994) acts as a good approximation in a sub-canopy, can also be called into question. At our site the mean difference between 3-D and 1-D rotated fluxes was negligible (0.5%–3.8%).

In calm conditions typically that of nighttime, CO_2 is stored in the air below the measurement level and the flux measured by EC can not be interpreted as the balance between respiratory and assimilatory fluxes from all of the ecosystem components below the measurement level (Goulden *et al.* 1996, Aubinet *et al.* 2000, Falge *et al.* 2001). There are two common ways to correct the fluxes for this phenomenon: (1) To calculate the change of the CO_2 storage every period and then add the storage flux to the EC flux (Markkanen *et al.* 2001), or (2) to reject measurements made under insufficient mixing and use data collected under similar but turbulent conditions to interpolate the missing observations using, for instance, the regression between temperature and the flux (Aubinet *et al.* 2000, Falge *et al.* 2001). The accurate determination of the storage term needs concentration measurements at least at two heights between the surface and the measurement level (Yang *et al.* 1999) which we had not at the forest floor. Instead, we calculated the storage flux assuming concentration changes at the measurement level to be representative for the whole air column between the forest floor and the measurement height, and find that the storage term was negligible when the σ_w-mixing criterion was used. Therefore we consider that diurnal changes (Fig. 7) and annual exchange of CO_2 (Fig. 10) are, to a great extent, free from storage effects.

If the storage effect is small, why does the absolute level of CO_2 flux measured by EC (F_c) differ from flux measured using the chamber technique (F_s) (Fig. 7a and b) particularly when the diurnal changes are similar and comparable? Our results suggest that the high-frequency CO_2-transport associated with small-scale eddies, which are either beyond the detection limit of eddy covariance setup or suspect to the high-frequency attenuation caused by the measurement setup, can induce at a maximum ~10% underestimation of the total exchange rate of CO_2. However, this underestimation can only partly explain the difference between the methods. One probable reason could be the great spatial variation in soil CO_2 efflux. Pumpanen *et al*. (2003) measured soil CO_2 effluxes from 10 collars installed randomly at the same site as in our study and found that the CV (coefficient of variation) in soil CO_2 effluxes ranged from 0.18 in the spring to 0.45 in July and August. The comparison of the fluxes measured with the automated chambers used in this study with the artificially generated known CO_2 effluxes showed that the fluxes measured with the chambers were overestimated by on average 8% (Pumpanen *et al*. 2004). The systematic error resulting from the chamber method has been corrected from the flux values presented in this study, thus the difference observed between the two methods probably originates from the spatial variation in soil CO_2 efflux or from the transport mechanisms not detectable by the eddy covariance method. The three automated chambers used in this study, were located on average about 40 m northeast from the eddy covariance tower which means that they were on the edge of the flux footprint area. However, the soil thickness and surface vegetation were about the same throughout the site.

Another probable reason may be the importance of horizontal or non-turbulent transport of CO_2. In general, it means that some of the respired CO_2 leaves the forest floor with mechanisms that can not be detected by eddy covariance. These possible transport phenomena may include: katabatic flow, dispersive flux related to local convection cells, slow diffusion or high-frequency fluctuations undetectable with eddy covariance technique (Aubinet *et al*. 2000). Theoretically a topographically-induced katabatic flow may increase the horizontal advection of CO_2 during stable nights. However, we have no evidence of this at present. Thus, more studies are needed to investigate this possibility. It is difficult to determine the systematical errors related to the eddy covariance and chamber methods. Norman *et al*. (1997) compared soil fluxes measured with closed dynamic system with that of eddy covariance, but the results were not consistent. On the other hand, Law *et al*. (1999) found good agreement between the methods.

We were able to make EC measurements for a full year under varying conditions. We detected annual and diurnal variability of the fluxes and got results which were generally in good agreement with results reported earlier from similar but campaign-wise studies. Our results obtained for daily CO_2 flux from the forest floor and its contribution to the ecosystem scale CO_2 exchange were comparable to the results reported earlier (Baldocci and Vogel 1996, Blanken *et al*. 1998, Constantin *et al*. 1999, Law *et al*. 1999) for a similar temperature range and type of forest. Vesala *et al*. (2005) concluded the annual forest floor photosynthesis below a Scots pine stand contributed about 10% of the gross photosynthetic production of the whole stand, i.e. about 100 g(C) m^{-2} a^{-1}. Thus, when this was added to the annual net exchange between the forest floor and the atmosphere (~320 g(C) m^{-2} a^{-1}), the annual forest floor respiration would be roughly 400 g(C) m^{-2} a^{-1}. This represents about half of the total ecosystem respiration of 813–894 g(C) m^{-2} a^{-1} derived from the above-canopy measurements (Kolari *et al*. 2004). For the reasons and uncertainties discussed above, the annual net exchange of carbon between the forest floor and the overlying atmosphere (Fig. 10) should be interpreted qualitatively rather than being simply a quantitative and precise result of the actual exchange. It should also be noted that direct comparison of below and above-canopy eddy covariance measurements is not accurate because the flux footprint area increases with increasing measurement height and possible spatial heterogeneities in surface characteristics can induce errors.

The diurnal variation in trunk-space CO_2-flux was significant during the summer. During this season, the midday depression of the CO_2 flux observed was mainly caused by the photosynthetical CO_2 uptake by the forest floor vegetation. The forest floor photosynthesis decreased F_c by 1.0–1.5 μmol m^{-2} s^{-1} relative to nocturnal values (Figs. 7b and 8). This CO_2-uptake rate agrees well with the values reported by Baldocci and Vogel (1996), Constantin *et al*. (1999) and Law *et al*. (1999) who measured below a boreal coniferous forest canopy uptake rates between 1.0–2.0 μmol m^{-2} s^{-1}. It is also consistent with the results from automated chamber measurements (Fig. 7a) and agrees reasonably well with the results of the manual chamber measurements of photosynthesis rate of forest floor vegetation which was little smaller than 3 μmol m^{-2} s^{-1} in July and August (data not shown). Annual variation in trunk-space F_c was mainly driven by changes in soil temperature (Fig. 9) and soil moisture content which regulates the soil CO_2 efflux (Pumpanen *et al*. 2003).

Conclusions

Our analysis of a full-year of trunk-space flux measurements suggests that the eddy covariance technique can be used to make continuous flux measurements below a relatively sparse boreal pine forest canopy when the pre-conditions of the method, which include stationarity, horizontal homogeneity and sufficient turbulent mixing, are met. Trunk-space fluxes in a Scots pine canopy were driven by intermittent and strong turbulent events associated with large eddies (size 15–100 m) that penetrate the whole canopy and are relatively easy to detect with the EC method. However, since we were applying the method under complex conditions, normal quality analysis methods could not be used. We found it preferable to use the standard deviation of the vertical velocity as a measure of the turbulent mixing in trunk-space instead of the friction velocity threshold.

Turbulent fluxes from the forest floor are a significant part of the ecosystem scale fluxes of sensible heat, latent heat and CO_2 in relatively open-canopied forests. In a boreal Scots pine forest, the forest floor contribution was 20%–40%. Sensible and latent heat fluxes were suspect to a clear annual and diurnal course driven by changes in environmental variables such as temperature, stability of the surface layer and available energy. The forest floor was a source of carbon all-year-round but the strength of the source varied annually and diurnally. The annual variation primarily followed the changes in soil temperature whereas the diurnal course, during the growing season was influenced by understory photosynthesis. Photosynthetical uptake of CO_2 by the forest floor vegetation decreased daytime fluxes by 1.0–1.5 μmol m^{-2} s^{-1} relative to nocturnal fluxes, but the uptake was not enough to compensate for the respirative carbon losses.

Our eddy covariance results of sensible heat and latent heat exchange as well as CO_2-exchange are comparable to those reported earlier for similar types of boreal forests. The diurnal variation in CO_2 flux was in good agreement with chamber measurements but there was a systematic difference in the absolute level of the fluxes. Trunk-space eddy covariance measurements provide a spatially averaged estimate of the heat and mass exchange between the soil/understory and the atmosphere and can, especially, give useful information on temporal changes in the fluxes. Used together with chamber techniques, the eddy covariance method can provide a useful tool to examine the roles of different parts of a forest ecosystem in forest-scale energy and carbon balance.

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