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ARTICLE

An Attempt to Close the Daytime Surface Energy Balance Using Spatially-Averaged Flux Measurements

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Abstract Single-tower eddy-covariance measurements represent the complete surface flux of a scalar only under idealized conditions. Therefore, we often find an underestimation of energy fluxes expressed as a lack of energy balance closure at many sites. In this study, a multi-tower approach to measure atmospheric energy fluxes based on spatial averaging is evaluated and possible mechanisms causing a lack of energy balance closure are analysed, focussing on daytime data only. It is shown that the multi-tower technique is also unable to measure the entire flux for our site, likely because the assumption of horizontal homogeneity is violated. Heterogeneity-induced and buoyancy-driven quasi-stationary circulations are probably the dominant processes causing the underestimation of energy fluxes. A dependence of the energy balance residual on stability is found, with residuals close to zero for stable stratification, a maximum under unstable to near-neutral conditions and still relatively large residuals for stronger instability. Assuming the processes transporting energy and CO_2 are similar, the implications on long-term CO_2 flux measurements are analysed. Accordingly, the resulting selective systematic error of cumulative net ecosystem exchange estimates for agricultural regions such as ours can be of the order of more than 100%, since mainly the fluxes during periods of net CO_2 uptake are underestimated while periods of net CO_2 release are much less affected by this bias. Further investigations about this issue are highly warranted.

Keywords Eddy covariance · Energy balance closure · Latent heat flux · Net ecosystem exchange · Sensible heat flux

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1 Introduction

The eddy-covariance technique is the most direct method for quantifying the turbulent exchange of energy and trace gases between the Earth's surface and the atmosphere. However, this method is based on several assumptions, such as horizontal homogeneity and zero mean vertical wind velocity. In addition, horizontal and vertical advection and horizontal and vertical flux divergence are not captured. Among other issues, this can lead to a bias of the flux estimate that becomes apparent in a lack of energy balance closure, usually ranging between 10 and 30% of the available energy at the surface (Desjardins 1985; Lee and Black 1993; Twine et al. 2000; Wilson et al. 2002; Culf et al. 2004; Oncley et al. 2007; Foken 2008). Several large-eddy simulations (LESs) support the thesis that single-tower eddy-covariance measurements using temporal averaging cannot completely represent the true surface flux (Kanda et al. 2004; Inagaki et al. 2006; Steinfeld et al. 2007; e.g. Huang et al. 2008; Wang 2010). According to these studies, the resulting flux bias depends on several factors, such as wind speed, measurement height, averaging time, degree of horizontal heterogeneity and atmospheric stability. Experimental studies also found correlations of these factors with the magnitude of the lack of energy balance closure (Mauder and Foken 2006; Foken et al. 2006; Mauder et al. 2007b). Since both energy and matter are transported by the same atmospheric motions it is very likely that eddy-covariance measurements of CO_2 fluxes in large-scale networks are also biased. Sometimes a correction of these CO_2 flux estimates according to the lack of energy balance closure is advocated (Twine et al. 2000), although there is little evidence that the relative flux bias is the same for different scalars. On the contrary, experimental results suggest that scalar similarity cannot generally be assumed for large-scale transport processes, because source/sink distributions can be different in heterogeneous terrain, and active and passive scalars show different behaviour (Ruppert et al. 2006; Mauder et al. 2007a; Guo et al. 2009). Since the common lack of energy balance closure is mostly found during daytime, we can then assume that, predominantly, net CO_2 uptake values are affected by the processes behind this flux bias, rather than net CO_2 release estimates obtained at nighttime. This then leads to a selective systematic error (Goulden et al. 1996) for monthly and annual estimates of net ecosystem exchange (NEE).

Large-eddy simulation results show that the entire flux can be captured if spatial instead of, or in addition to, temporal averaging is applied when calculating the covariance (Steinfeld et al. 2007). This principle can either be applied using an airborne platform or using a multi-tower set-up, which allows continuous measurements. The analysis of Mauder et al. (2007a) showed that it is possible to measure the entire range of transporting atmospheric motions using aircraft measurements with a flight length of 120 km, and that mesoscale flux contributions at wavelengths larger than 2 km can explain the lack of energy balance closure measured at single-tower sites in the vicinity of the flight track.

The study of Mauder et al. (2008b, henceforth M08) attempted to determine the sensible heat flux based on a spatial average of 25 slow-response temperature sensor towers and a fast-response measurement system at one site in the centre of this set-up. As derived in M08, the sensible heat flux H_{TS} can then be calculated by time-spatial averaging as

$$H_{TS} = w(T - [T]), \tag{1}$$

where w is the vertical wind velocity, T is the air temperature, [ ] stands for a time-planar average and the overbar denotes a temporal averaging operator. This approach, which is based on the continuity equation, still invokes stationarity and planar homogeneous flow, but allows for local subsidence to exist while it assumes area-averaged subsidence to be negligible. The experiment presented in M08 was carried out over flat farmland in 2007, and additional contributions to the sensible heat flux of more than 50 W m^{-2} were found by applying the spatial eddy-covariance method as opposed to the standard temporal eddy-covariance method. This is approximately the same magnitude as the commonly found energy balance residual. The objective of the present study is to apply this same approach for the measurement of the latent heat flux and also measure the ground heat flux and net radiation in order to assess the energy balance closure. According to M08, Eq. 1 can be Reynolds-decomposed in

$$H_{TS} = \overline{w}(\overline{T} - [T]) + \overline{w'T'}, \tag{2}$$

which has the advantage that the time-averaged eddy-covariance based flux at a single location (second term on the right-hand-side (rhs) of Eq. 2) can be determined independently from the vertical mass transfer or convective flux (first term on the rhs of Eq. 2). This reduces the need for exact inter-calibration of the fast response scalar measurements with the slow response spatial network. We applied this same approach for the latent heat flux λE_{TS}, substituting the specific humidity q for T in Eq. 2,

$$\lambda E_{TS} = \overline{w}\,(\overline{q} - [q]) + \overline{w'q'}. \tag{3}$$

2 Inter-Comparison Pre-Experiment

In order to ensure the quality of the measurements, a pre-experiment inter-comparison was carried out over short grass at the central experimental farm of Agriculture and Agri-Food Canada in Ottawa, Ontario, Canada (45°23′23″ N, 75°42′53″ W, 81 m a.s.l.).

2.1 Ultrasonic Anemometers/Thermometers

Two ultrasonic anemometers/thermometers were used for this study, one CSAT3 (Campbell Scientific Inc. (CSI), Logan, Utah) and one Kaijo Denki model DAT-310, probe TR-62TX, both having an asymmetrical design, thereby minimizing flow distortion except for a small sector from behind, with sonic pathlengths of 0.12 and 0.10 m, respectively. CSAT3 sonics have been well-tested in several previous inter-comparison studies (Foken and Oncley 1995; Mauder et al. 2006, 2007c), and therefore such an instrument was appropriate for use as a reference. In order to check the comparability of the Kaijo Denki, these two sensors were operated side-by-side during the period from 25 April 2008 1400 EST to 14 May 2008 0700 EST (Eastern Standard Time). The canopy was short grass, and the measurement height was the same as for the main experiment, 2.6 m. The three variables that are most relevant for the calculation of heat fluxes, i.e. σ_w^2, $\sigma_{T_s}^2$ and $\overline{w'T_s'}$, show a reasonably good agreement without any correction being necessary. The root-mean-square differences (RMSD) for these variables were 0.020 m^2 s^{-2}, 0.045 K^2 and 0.0087 K m s^{-1}, respectively.

2.2 Slow Response Temperature and Humidity Sensors

Temperature and humidity were measured at 25 locations using HOBO 12-bit smart sensors (Onset Computer Corporation, Bourne, Massachusetts), part # S-TMB-M002 for temperature and part # S-THB-M002 for humidity, in combination with naturally ventilated HOBO solar radiation shields (Onset, part # M-RSA). The accuracy and precision of the temperature measurement using these shields is addressed by Mauder et al. (2008a). Data were recorded at each tower using HOBO Micro Station data loggers (Onset, part # H21-002) at a sampling interval of 30 s, with all sensors mounted at a height of 2.6 m. For an inter-comparison pre-experiment, the 25 towers were set-up in a 5 by 5 grid covering an area of approximately 10 m by 10 m, similar to Mauder et al. (2008a). Data from this set-up were analyzed for the period from 24 April 2008 1600 EST to 01 May 2008 1559 EST, and a total of 20,160 30-s observations were available for the comparison. As mentioned above, when the flux is calculated using Eq. 2 there is no need for an exact inter-calibration between the fast-response and slow-response sensors. Only the measurements of the 25 HOBO sensors need to be in reasonably good agreement. The sensors of one tower (L11) were arbitrarily selected as a reference system in order to calculate the comparability/RMSD for all other 24 systems. For the temperature measurements, the maximum RMSD was 0.130 K, the average RMSD was 0.068 K and the median RMSD was 0.067 K. For humidity, the maximum RMSD from L11 was 0.259 g m^{-3}, the average RMSD was 0.096 g m^{-3} and the median RMSD was 0.073 g m^{-3}. Four sensors showed a particularly large deviation from the reference L11, i.e. L8, L17, L21, L24, most of it originating from nighttime data. No plausible explanation could be found for this behaviour. Since these four sensors always agreed well between each other, we decided to correct their measurements during the main experiment, using the difference between L24 and L19, which were operated side-by side at site 20/25.

3 Experimental Set-Up

The observation area of the main experiment was the same as for the study presented in M08, an agricultural area with the dimensions of approximately 4 km × 4 km located in south-west Ottawa, Ontario, Canada (45°18′09″ N, 75°46′00″ W, 88 m a.s.l.). The measurement period was from 27 June 2008 1200 EST to 6 August 2008 0900 EST. The locations of the 25 slow response HOBO temperature/humidity systems were the same as in M08. The crop types cultivated in the observation area were forage, corn, soybean and wheat, and in addition, there was one field with smaller test plots for various species and some fallow land. The south-western part of the area was covered with grass, shrubs and small trees (Fig. 1). An observation height of 2.6 m above ground level was chosen, since it is a typical height for measurements over agricultural land where large energy imbalances have been reported (Mauder et al. 2006; Foken 2008). According to the LES of Steinfeld et al. (2007), a number of 25 slow-response sensors is sufficient to obtain a representative spatial average. The temperature sensors were distributed in a regular 5 × 5 grid covering an area of 3.5 km × 3.5 km, with the grid slightly adapted to the local circumstances due to accessibility. All sensors were equally weighted for calculating the spatial average.

Fast-response measurements were carried out at the central site 13 over grassland for calculating the regular temporal eddy-covariance flux of sensible heat, water vapour and CO_2. This system was equipped with a CSI CSAT3 sonic anemometer/thermometer and a LI-7000 closed-path infrared gas analyser (LiCor Biosciences Inc., Lincoln, Nebraska). Hence, this set-up together with the slow-response system was able to measure H and λE according to Eqs. 2 and 3. All fast-response data were recorded on a Campbell CR23X data logger at 20 Hz, and temporal averaging and the calculation of covariances was done during post-processing for 30-min intervals. Webb et al. (1980) corrections for temperature and water vapour density effects were not made because temperature fluctuations are eliminated using a closed path analyser and the water vapour density correction was small. An extension of the averaging time did not lead to significantly higher fluxes, as has been observed for other sites (e.g. Finnigan et al. 2003; Mauder and Foken 2006). Since the system at site 13 relied

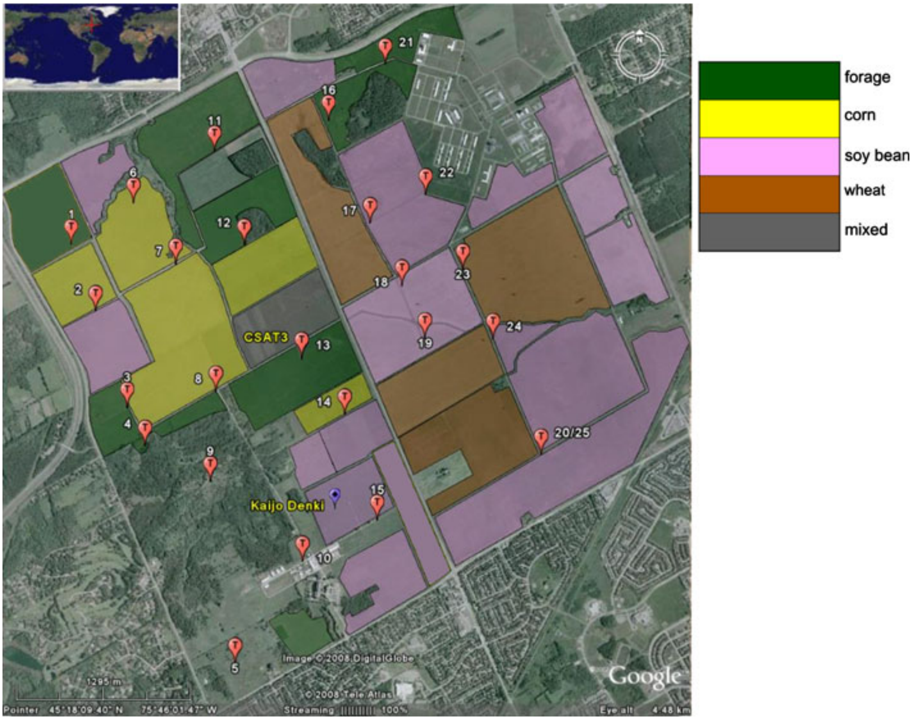


Fig. 1 Overview of the measurement set-up with land use information. The locations of the 25 HOBO temperature/humidity sensors are indicated by red pins labelled with a T and the site number. This image covers an area of approximately 5 km by 5 km

on solar power and batteries the LI-7000 was shut off at night between 1900 and 0700 EST in order to reduce power consumption.

From the results of M08 (M08, Fig. 10), we expected a consistent mean downward vertical motion at least during some of the days at site 13 and a correlated mean upward motion on the soybean field 1.2 km south of this central site, which was essentially bare soil or only sparsely vegetated during the first two weeks of the observation period. The Kaijo Denki sonic system, which had been inter-compared with the CSI CSAT3 previously (see Sect. 2.1), was set-up on this field (Fig. 1). These measurements were sampled at 10 Hz through a Campbell CR21 data logger that was connected to a personal computer where the data were stored. In order to compensate for a potential misalignment of the sonics, the wind measurements were transformed into a coordinate system parallel to the mean streamline using the planar fit method (Wilczak et al. 2001), however with a small modification compared to the original method. Since it was our objective to determine the actual non-zero mean velocity we did not use the b_0 offset from the multiple linear regression analysis for this correction but instead subtracted a sensor-specific w bias. This w bias was determined by placing the sonics in a closed box in the laboratory and recording the wind signal over several minutes. The result was an offset of 0.026 m s^{-1}, which was generally subtracted from the measured w values.

At site 13, in addition to the atmospheric energy fluxes, all other relevant terms of the surface energy balance were measured. Net radiation R_n was quantified at 2 m height using a Schulze-Däke net radiometer (Dr. Bruno Lange GmbH, Düsseldorf, Germany), which has an uncertainty below $\pm 20\,\mathrm{W\,m^{-2}}$ according to Kohsiek et al. (2007). The soil heat flux at a depth of 0.08 m was measured using two heat-flux plates (CAPTEC, Lille, France), while the heat storage in the soil layer above was determined using two model 107 temperature probes from Campbell Scientific Inc. at a depth of 0.02 and 0.06 m. Based on the sensitivity analysis of Liebethal et al. (2005) and the QA/QC results of Mauder et al. (2006), we estimate the accuracy of this combined approach for measuring the ground heat flux at the surface G to be within $\pm 15\%$ but at least $\pm 15\,\mathrm{W\,m^{-2}}$. Since the observations were made over low vegetation and the measurement height of the atmospheric fluxes was only 2.6 m, we assumed the changes of the heat and biomass storage in the canopy space to be negligible. The canopy height at site 13 ranged between 0.30 m and 0.75 m. In comparison, Oncley et al. (2007) showed for a 1-m high cotton crop in California that both heat and biomass storage in the canopy were less than $10\,\mathrm{W\,m^{-2}}$.

4 Results

The sum of the sensible and latent heat fluxes from the single-tower eddy-covariance measurements and from spatial averaging are plotted versus the available energy at the surface ($-R_n - G$) in Fig. 2. The lack of energy balance closure remains almost the same as indicated by the regression equations shown in Fig. 2. This was unexpected since M08 found sensible heat fluxes H either higher than or equal to the pure eddy-covariance flux when the convective contribution was included. In order to explain this finding we plotted the time series of H and λE in Fig. 4 for a period of four and a half days with relatively high incoming solar radiation. Significant convective fluxes of both sensible and latent heat did occur, larger than $30\,\mathrm{W\,m^{-2}}$ at times. Overall the magnitude of these fluxes is smaller than the observed residual. Moreover, the direction of these fluxes was often opposite and of the same magnitude so that they cancel each other out when the sum of both fluxes is calculated. The mean diurnal course of the energy balance residual is shown in Fig. 3. No latent flux data were available before 0700 EST and after 1900 EST, because the LI-7000 and its pump were switched off at night, and therefore also no energy balance residual could be calculated during this time period. The residual was largest during midday with values over $100\,\mathrm{W\,m^{-2}}$ on average from 0830 to 1400 EST. In the evenings, around sunset, the residual went down to zero.

According to Eqs. 2 and 3, a non-zero vertical wind is a precondition for additional vertical fluxes in addition to the regular eddy-covariance flux. The mean diurnal course of the vertical wind velocity $\overline{w}$ for the central grassland site number 13 is shown in Fig. 5 after transformation of the coordinate system parallel to the mean streamlines, subtraction of the w bias and filtering for rainy periods. The maximum $\overline{w}$ was $0.01\,\mathrm{m\,s^{-1}}$ and the minimum was $-0.08\,\mathrm{m\,s^{-1}}$. The majority of the data (98%) were below zero indicating a downward mean mass transfer. The mean vertical velocity was consistently below zero during the daytime with a minimum of $-0.036 \pm 0.011\,\mathrm{m\,s^{-1}}$ for the period between 1330 and 1400 EST. During nighttime, the mean $\overline{w}$ was slightly negative or close to zero. Combining the results of Figs. 4 and 5, the convective energy fluxes with opposite signs can be explained when $\overline{w} < 0$ and $\overline{T} < [T]$ and $\overline{q} > [q]$, which leads to $H_{EC} < H_{TS}$ and $\lambda E_{EC} > \lambda E_{TS}$, so that the energy balance residual is hardly altered. The histograms presented in Figs. 6, 7, and 8 show indeed that this combination of conditions was prevalent with $\overline{w}$ and $\overline{T} - [T]$ being negative most of the time and predominantly positive values for $\overline{q} - [q]$. Although we have spatial information about the temperature and humidity fields and wind measurements, it is not possible to calculate meaningful advection estimates from these data. The distance between the slow response sensors is too large (700–1,000 m) to assume a linear scalar gradient in

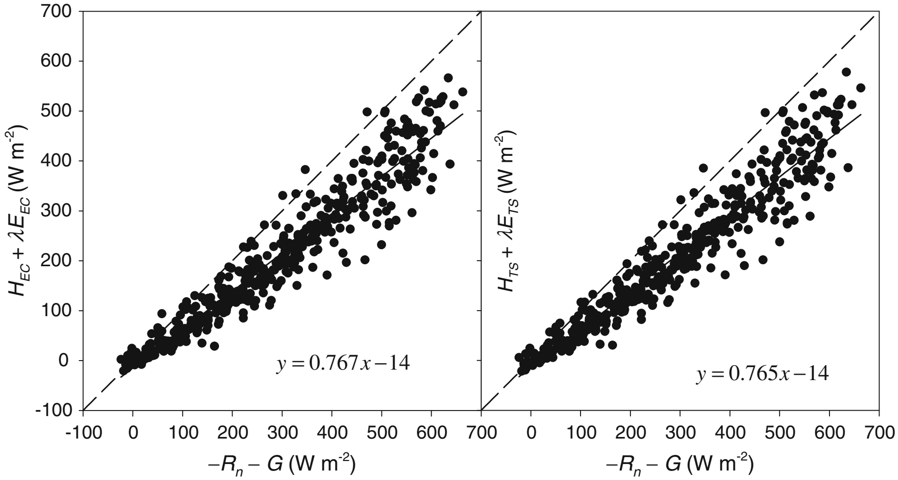


Fig. 2 Energy balance scatter plots with linear regression for the pure turbulent fluxes (*left panel*) with 30-min averaging time and the atmospheric fluxes including the convective contribution (*right panel*); all terms positive for fluxes away from the surface

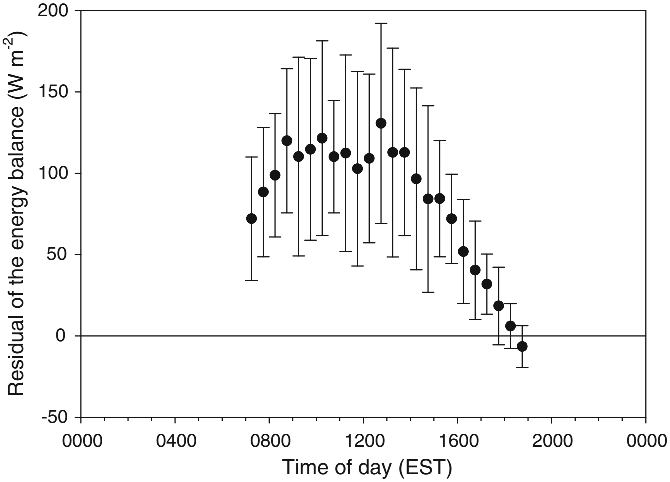


Fig. 3 Mean energy balance residual at site 13 as a function of the time of day including the standard deviation for the entire observation period from 27 June to 6 August 2008. A positive residual implies that the available energy is larger than the sum of the turbulent fluxes

between, which would be a precondition. If we were to calculate horizontal advection on this basis (Eq. 4) the resulting fluxes would only be on the order of a few W m^{-2}, using

$$H_{h\_adv} = \int_0^{z_m} \left[\overline{u} \frac{\partial \overline{T}}{\partial x} + \overline{v} \frac{\partial \overline{T}}{\partial y} \right] dz. \quad (4)$$

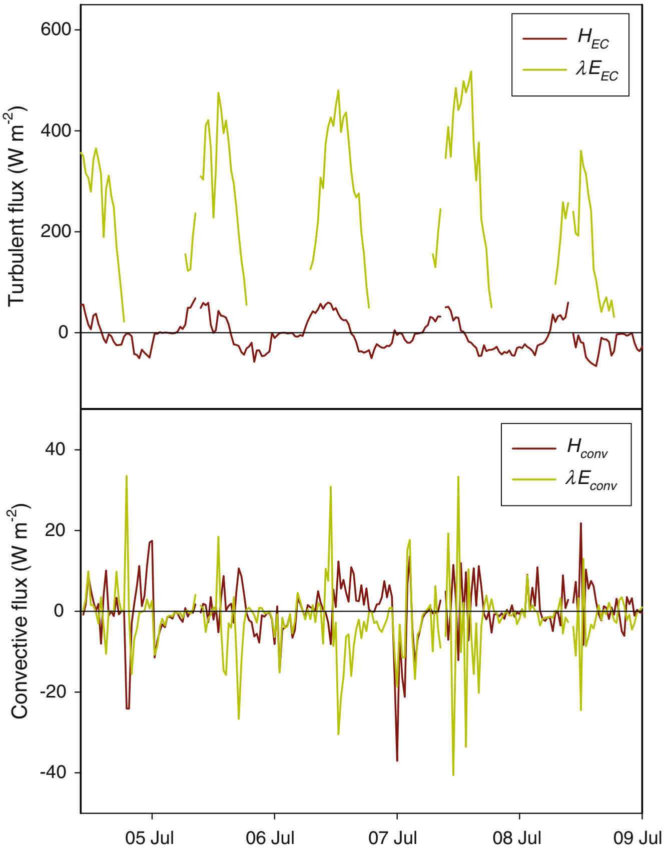


Fig. 4 Turbulent energy fluxes at site 13 as measured by the temporal eddy-covariance method (*top panel*) and convective fluxes (*bottom panel*) for the period from 4 July to 9 July 2008

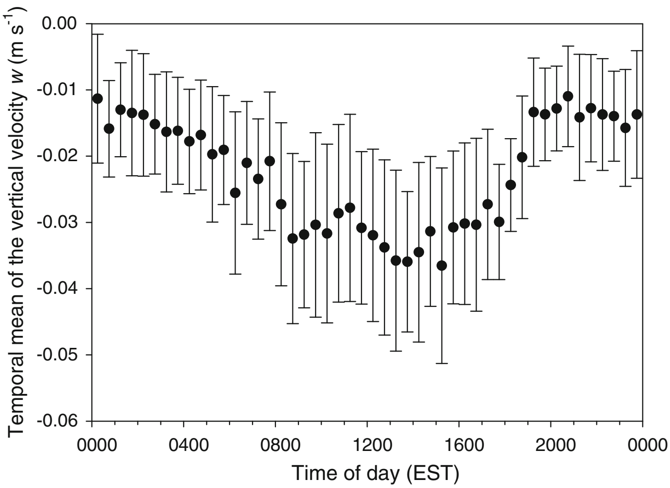


Fig. 5 Temporal mean of the vertical wind velocity at site 13 as a function of the time of day including the standard deviation for the entire observation period from 27 June to 6 August 2008

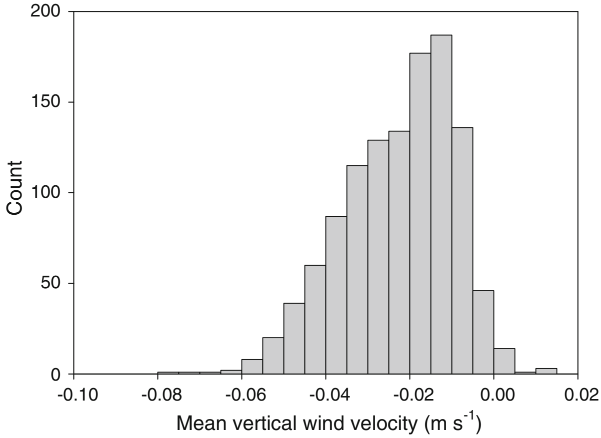


Fig. 6 Histogram of the mean vertical wind velocities $\overline{w}$ at site 13 for the observation period from 27 June to 6 August 2008

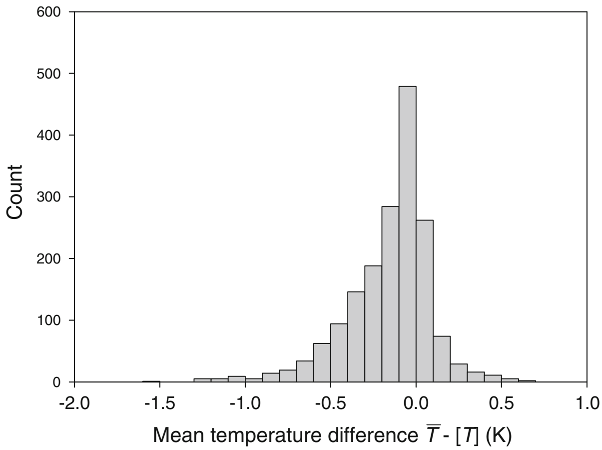


Fig. 7 Histogram of the difference between the local time-averaged temperature at site 13 and the time-spatial temperature average $\overline{T} - [T]$ for the observation period from 27 June to 6 August 2008

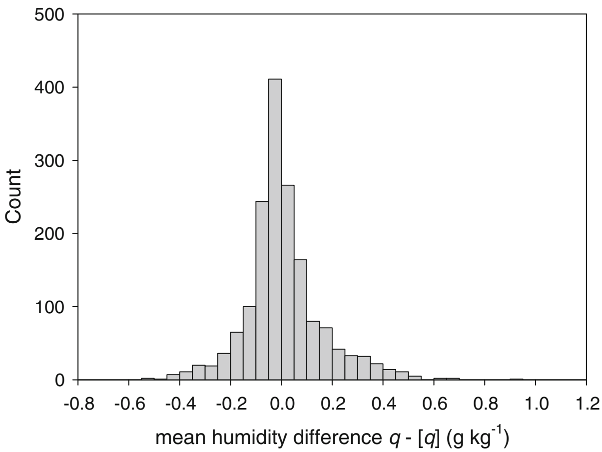


Fig. 8 Histogram of difference between the local time-averaged specific humidity at site 13 and the time-spatial specific humidity average $\overline{q} - [q]$ for the observation period from 27 June to 6 August 2008

5 Discussion

5.1 Validity of the Assumptions for the Measurement Approach

The results presented in Sect. 4 show that the method presented in M08 based on spatial averaging of a slow-response multi-tower set-up is not able the measure the entire energy exchange between the earth's surface and the atmosphere. Considering the assumptions made for the derivation of this approach can help us to understand its deficiencies. These are: (I) stationarity, (II) planar homogeneous flow, (III) negligible area-averaged subsidence, and (IV) dispersive heat fluxes are small. Stationarity was evaluated using the test of Foken and Wichura (1996). Accordingly, non-stationarities occurred sometimes, in particular during transition periods in the morning and evenings and at nighttime, but did not constitute a general problem during daytime. The assumption of planar homogeneous flow is hard to verify with our set-up: the grassland area around the central site was completely flat and had a size of 780 m $\times$ 410 m; the main wind direction was south-westerly, where the fetch was 350 m long, with the vast majority of the data (80%) between directions 110° and 300°. According to Jegede and Foken (1999), using a measurement height of 2.6 m minus a zero-plane displacement height of 0.2 m results in a minimum homogeneous fetch requirement of 64 m for near-neutral and unstable conditions. This criterion, which is applicable for general single-point eddy-covariance measurements, was amply fulfilled for our set-up since the distance to the next land cover change was greater than 100 m for all directions. However, at a larger scale planar homogeneous flow was probably violated since the grassland area was surrounded by fields with different crops and the farm area in turn was surrounded by streets and houses of the city of Ottawa. An indication of a large-scale circulation between the central site and a soybean field, 1.2 km south of it, can be seen in Fig. 9. During the first part of the experiment, from 27 June to 6 July 2008, when the soybean crop was relatively short (<0.15 m) and the soil was not completely covered, $\overline{w}$ was consistently positive there while it was negative over the more transpiring and cooler grassland area. As soon as the soybean grew higher, after 7 July, $\overline{w}$ was still negative at the grassland site but scattered around zero at the soybean site with more negative values than positive ones. This indicates that the balancing upward motion must have occurred elsewhere, one possibility being the warmer city areas around the farm area for instance. The results shown in Fig. 9 also cause considerable doubts that assumption III, i.e. negligible area-averaged subsidence, is valid; Assumption IV cannot be evaluated from our measurements. However, based on other studies (Poggi et al. 2004) it seems likely that dispersive heat fluxes, which result from spatial correlations between the time-averaged velocity and scalar fields, are indeed relatively small over an observation area with flat terrain and low vegetation.

The fact that assumption II (planar homogeneous flow) is likely to be violated on relevant scales means that horizontal advection and horizontal flux divergence are very significant. The violation of the negligible area-averaged subsidence assumption means that one has to measure the vertical velocity also spatially, not only at the central site, if one wants to apply this approach of calculating a flux based time-spatial averaging (M08). It has to be noted that all these assumptions also underlie the conventional single-tower eddy-covariance method, and in addition, the single-tower eddy-covariance method also assumes local subsidence to be zero, which is clearly shown to be not true for our site.

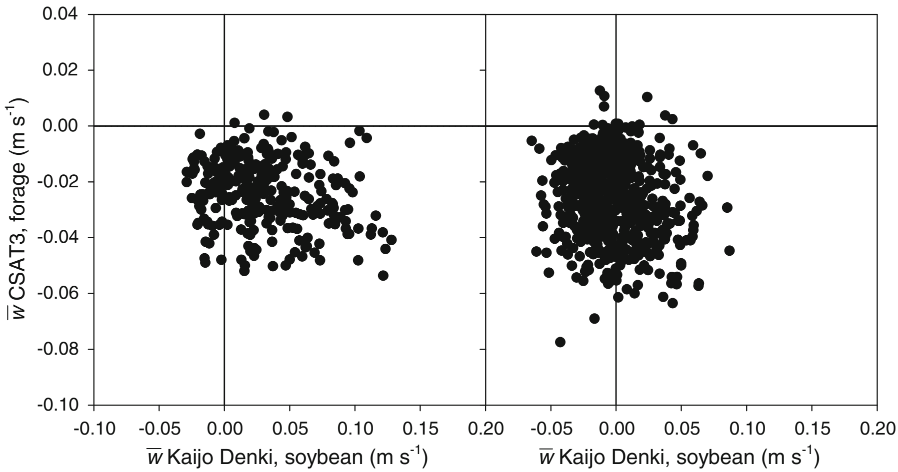


Fig. 9 Vertical wind velocities $\overline{w}$ at the central site measured over grass (CSAT3) versus $\overline{w}$ over bare soil/soybean (Kaijo Denki) for the period before (*left*) and after (*right*) 7 July 2008, when the soybean crop reached a height of 0.15 m

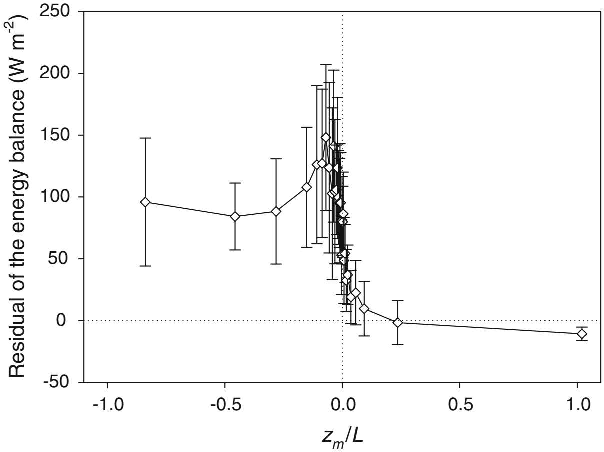
5.2 Effect of Temperature on the *w* Measurements

In a laboratory experiment, which was conducted in December 2009, we found a temperature dependence of the CSAT3's *w*-offset. At zero wind the sonic anemometer showed a linear relationship ($R^2 = 0.93$) between T_s and w, with a w-offset of 0.0 m s^{-1} for 4°C and +0.020 m s^{-1} for 20°C. The difference between these results and the +0.026 m s^{-1} at room temperature from the summer of 2008 can probably be explained by a small ageing effect of the transducers. Regarding the validity of the velocities shown in Fig. 5, we can state that the measured $\overline{w}$ values are in the opposite direction to the laboratory detected *w*-offset and the difference between daytime and nighttime velocities would be even larger than shown in this Figure, if a temperature-dependent correction was applied. Since there would be some uncertainties related to such a correction, e.g. the 1-year delay between the field measurements and the laboratory experiment, and using T_s as a proxy for the transducer temperature, we decided to choose a conservative approach, not applying any further correction to the *w* data. However, for future experiments of that kind we propose to measure the transducer temperature in the field and do a careful calibration of the temperature dependent *w*-offset in the laboratory before and after the field campaign.

5.3 Analysis of the Transport Mechanisms at Work

Quasi-stationary circulations (Finnigan 1999) and eddies with larger time scales than the covariance averaging time (Sakai et al. 2001; Finnigan et al. 2003) cannot be captured with single-tower measurements and are therefore likely candidates for causing the lack of energy balance closure. An extension of the averaging time is one possibility to include these otherwise neglected flux contributions. A comparison of the sensible and latent heat fluxes for averaging times of 240, 480 and 720 min with the corresponding arithmetic means of the 30-min fluxes showed no systematic differences regarding the magnitude of the flux estimates. Extending the averaging time had little effect on the atmospheric energy fluxes, less

Fig. 10 Binned energy balance residual, based on the 30-min eddy covariance estimates, including standard deviation, versus the stability parameter z_m/L measured at site 13 for the observation period from 27 June to 6 August 2008. Only daytime (0700 to 1900 EST) data are presented



than $11\,\mathrm{W\,m^{-2}}$ on average. This means eddies with very large time scales can be excluded as a potential transport mechanism. Hence, only quasi-stationary circulations remain as a possibility, probably caused by surface heterogeneity and buoyancy effects. At least for the period from 27 June to 6 July, we are able to identify the newly planted soybean field, where the Kaijo Denki sonic was located, as the upward part of this circulation, which balanced the negative mean vertical mass flow over site 13 (Figs. 5, 9). The corresponding downward part of this circulation with a mean downward motion was measured at site 13 over the cooler and readily transpiring grassland. The sensible heat flux at noon for this period was $98\,\mathrm{W\,m^{-2}}$ on average over the newly planted soybean field and only $46\,\mathrm{W\,m^{-2}}$ at the central site 13.

If the lack of energy balance closure can be attributed to some kind of convective flux, one would expect a relation between the energy balance residual and the atmospheric stability parameter z_m/L, where L is the Obukhov length. Stoy et al. (2006) found for their site that the residual increases with increasing instability, and Barr et al. (2006) report that the daytime energy closure fraction was highest for $-1 < z/L - 0.01$ for mature forests. For our measurements (Fig. 10), we found that the residual was close to zero during stable stratification, increased sharply for neutral conditions, reached a maximum of almost $150\,\mathrm{W\,m^{-2}}$ for slightly unstable stratification at around $z_m/L = -0.05$, and then decreased again towards values around $100\,\mathrm{W\,m^{-2}}$ for increasing instability. It should be mentioned again that only daytime data, from 0700 to 1900 EST, were analysed here. Hence, we speculate that the transport by heterogeneity-induced stationary circulations that cause the lack of energy balance closure do not develop during periods of stable stratification, are strongest during periods of near-neutral to slightly unstable stratification, and are still present during very unstable periods but not as pronounced.

We can exclude a number of other potential reasons for a lack of energy balance closure through this plot (Fig. 10). If turbulent transport is very little during periods of stable stratification, the other terms R_n and G do not have to be zero at all. Energy balance closure approaches zero only when R_n and G balance each other. The fact that this is the case, gives us confidence in our R_n and G measurements. Furthermore, energy balance closure could be different from zero for stable stratification if there was a topography driven katabatic advection. This cannot be seen in our data either. With increasing stable stratification, flux footprints extend much longer, and since for stable stratification absolute energy balance closure is close to zero, mismatch of flux footprints could be excluded.

Partitioning of the energy balance residual into the sensible and the latent heat fluxes could not be determined for this experiment. The studies of Ruppert et al. (2006) and Mauder et al. (2007a) show that similarity between these two atmospheric energy fluxes cannot be assumed for the large-scale processes that contribute to the lack of energy balance closure. Stoy et al. (2006) argue that the top of the boundary layer will be relatively dry at times (which would decrease the measured λE), but it will almost certainly be colder than the surface layer. Accordingly, it is expected that eddy-covariance-measured H departs more from the true surface flux than λE if convection explains the bulk of the energy balance residual as hypothesised. This remains to be confirmed by direct measurements.

One might ask why covariances based on spatial averaging seem to provide the total flux for aircraft measurements (Mauder et al. 2007a) but the computation of the flux using a time-spatial average from 25 towers does not reduce the energy balance residual. The reason is that the assumption of horizontal homogeneity is violated for the tower measurements (see Sect. 5.1), whereas the natural mixing of eddies with increasing height leads to a blending of surface heterogeneities. Since aircraft measurements are usually conducted at higher altitudes, this problem is largely reduced.

5.4 Implications for CO_2 Flux Measurements

There are also problems with the eddy-covariance technique during nighttime due to stable stratification and low turbulence intensities, but these are not an objective of our study since measurements from these periods are usually rejected and a gap-filling model is applied to compute an NEE estimate (Falge et al. 2001). The focus of our study lies in daytime unstable or near-neutral conditions.

Some measurements have indicated that the similarity between the CO_2 and the water vapour flux is closer than between the sensible and latent heat fluxes for the processes that cannot be measured by the single-point eddy covariance (Twine et al. 2000; Mauder et al. 2007a). This makes sense since the distribution of sources and sinks is relatively similar for CO_2 and H_2O. Consequently, if the latent heat flux is underestimated we have to consider the possibility that the CO_2 flux eddy-covariance measurement is affected in a similar way. We do not know the partitioning of the energy balance residual between H and λE, but we showed that the lack of energy balance closure is restricted to daytime periods with near-neutral and unstable stratification. In Fig. 11, the effect of a daytime flux bias on the cumulative NEE estimate is simulated for CO_2 flux measurements that were also conducted on the soybean field where the Kaijo Denki sonic was located (see Fig. 1). This set-up comprised a Solent HS sonic anemometer (Gill Instruments Ltd., Lymington, UK) mounted at 3 m above the ground on a rotating platform, for automating alignment to the horizontal wind direction, and a LI-7000 gas analyser (LiCor Biosciences Inc., Lincoln, Nebraska) kept in a temperature controlled enclosure. The description of the closed-path IRGA set-up, flux correction and screening are provided in Pattey et al. (2006). Figure 11 shows the quality controlled and gap-filled data for the crop growing period of 2008 (i.e. between seeding and harvest), noting that we did not gap-fill individual 30- min nighttime data but rather the missing nights (Pattey et al. 2002). Through the screening of entire nights for calm conditions, the procedure consists in retaining an acceptable proportion of calm periods (usually lower or equal to 40%) on predominantly windy nights. Daytime data are gap-filled at the 30- min interval. The cumulative NEE as measured by single-tower eddy covariance was $-98\,\mathrm{g\,CO_2\,m^{-2}}$, and if we assume a 10% underestimation during daytime (from 0700 to 1900 EST) this would lead to a NEE of $-211\,\mathrm{g\,CO_2\,m^{-2}}$. A 25% underestimation, which is approximately equal

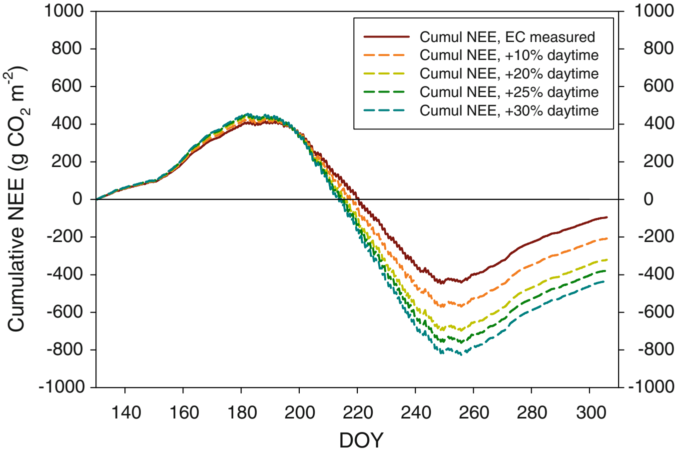


Fig. 11 Cumulative NEE measured over the soybean site, where the Kaijo Denki sonic was also located, for the growing period from DOY 130, 9 May, to DOY 305, 31 October, 2009. The *dashed lines* indicate the cumulative NEE if we assume a daytime flux bias by a certain percentage

to the energy balance residual that we found, would lead to a NEE of $-382\,\mathrm{g\,CO_2\,m^{-2}}$, and would correspond to a selective systematic error of almost 400%.

If the soil carbon pool is constant on an annual basis, the cumulative NEE for one year should be comparable with the yield estimate of the soybean crop. However, the validity of this assumption is not very certain. The crop rotation on this field was wheat in 2005, corn in 2006, wheat in 2007, and soybean in 2008. Residues may have a different rate of decomposition, and no meaningful direct C-pool measurements are possible within three years, as differences are small, but they do contribute to C change and sequestration. Despite this uncertainty we believe that a comparison with the yield estimate is interesting and can give some indication about the direction and the magnitude of the CO_2-flux bias. Yield was measured at harvest using a yield monitor (GreenStar Combine Yield Systems, Deere and Co., Moline, Illinois), which was back-calibrated regularly using weigh wagon load masses. The average dry soybean yield in 2008 (i.e., $200\,\mathrm{g\,DM\,m^{-2}}$) was equivalent to a net CO_2 soybean uptake of $320\,\mathrm{g\,CO_2\,m^{-2}}$, using a conversion factor of $0.62\,\mathrm{g\,DM\,g^{-1}\,CO_2}$. The conversion factor was calculated using the following methodology described in Pattey et al. (2001). The conversion of daily CO_2 canopy uptake into plant biomass equals the ratio between the linear rate of biomass accumulation derived from destructive sampling, divided by the corresponding rate of net daily CO_2 canopy uptake accumulation determined by eddy covariance. This estimate of $320\,\mathrm{g\,CO_2\,m^{-2}}$ is almost the value that we obtained for a +20% daytime adjustment ($-321\,\mathrm{g\,CO_2\,m^{-2}}$), supporting the hypothesis of a selective systematic error due to the processes that contribute to a lack of energy balance closure. Although random errors of 30-min flux estimates can be of the same order of magnitude, they will mostly cancel each other out over a year. Thus, we can state in accordance with Schmid et al. (2003) that propagation of systematic errors is much more important than random errors for the uncertainty assessment of cumulative NEE estimates.

However, we do not propose to do such a correction for NEE estimates from eddy-covariance measurements in general, since the partitioning of the energy balance residual and the impact of these mechanisms on the CO_2 flux measurement is not sufficiently known. Moreover, analyses of the cospectra for λE and CO_2 show that similarity at low frequencies cannot be generally assumed (Scanlon and Albertson 2001), suggesting that errors in water vapour fluxes might not directly translate to errors in CO_2 fluxes. Hence, it is only intended here to demonstrate the potential magnitude of this error, which is most probably also site-dependent and probably varies with time. It could be larger or smaller for other flux measurement sites.

6 Conclusions

In our study we reassess the method presented in M08 based on time-spatial averaging of a scalar and measurement of the absolute vertical wind velocity at one central site. We show that this method is not able to measure the entire exchange of energy between the surface and the atmosphere. A non-zero $\overline{w}$ can be one indicator of processes that lead to a lack of energy balance closure, but a lack of energy balance closure can also be found for periods when $\overline{w} \approx 0$. The missing energy is most probably transported into the atmosphere in the form of sensible and latent heat, though the partitioning between these additional fluxes is unknown. For this site, eddies of time scales greater than 30 min could be excluded as a potential mechanism contributing to the lack of energy balance closure. It is likely that heterogeneity-induced quasi-stationary circulations are responsible for the additional transport. Since these circulations usually balance a gradient between relatively warm and moist air near the surface and relatively cool and dry air in upper levels during daytime, the resulting net transport of energy has to be upwards in general, regardless of the location. This explains the observed systematic energy imbalance in one direction, i.e. turbulent transport is generally lower than the available energy at the surface.

If CO_2 fluxes are determined by single-tower eddy-covariance measurements, it is very likely that these estimates are also affected by the same processes. Since these processes occur mainly during daytime, when plants are taking up CO_2, this leads to a selective systematic error in NEE computations that can potentially be as large as several 100%. However, the true systematic error of CO_2 flux measurements from eddy-covariance towers is unknown since scalar similarity cannot necessarily be assumed for these processes. The development of different measurement techniques in addition to, or alternatively to, the conventional temporal eddy-covariance method using single-tower measurements is warranted. Numerical simulations can also improve our understanding of all transporting atmospheric motions. Considering the importance of this issue, the development of innovative approaches to quantify the complete atmospheric flux of a scalar, and not only the turbulent part that can be captured using single-tower flux measurements, should be of highest priority for the micrometeorological community.

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