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Comparison between Different Methods of Measurement of Momentum and Sensible Heat Fluxes over Canopies

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Different methods of measurement of momentum and sensible heat flux densities are presented and compared above a grass covered fallow. The aerodynamic (AD) and eddy covariance (EC) methods are presented and compared for both momentum and sensible heat measurements. In addition, the temperature fluctuation (TF) method is compared to the HEC method for the sensible heat flux measurement. The AD and EC methods are in good agreement for the momentum flux measurements. For the sensible heat flux, the AD method is very sensible to temperature errors. So it is unusable during night and gives biased estimations during the day. The TF method gives only estimations of the sensible heat flux. It is in good agreement with the EC method during the day but diverges completely during night, being unable to discerning positive from negative fluxes. From the three methods, the EC method is the sole that allows to measure continuously both momentum and sensible heat flux but it requires a loud data treatment. We present in this paper the algorithm used for this treatment.
Keywords. Eddy covariance, temperature fluctuation, aerodynamic method, flux measurement.

Comparaison de différentes méthodes de mesure des densités de flux de quantité de mouvement et de chaleur sensible au-dessus d'un couvert végétal. Plusieurs méthodes de mesure des densités de flux de quantité de mouvement et de chaleur sensible sont présentées et comparées au-dessus d'une jachère herbeuse. Pour la quantité de mouvement, nous présentons et comparons les méthodes aérodynamique (AD) et par covariance de turbulence (CT). Pour la chaleur sensible, outre ces deux méthodes, nous présentons également la méthode par mesure des fluctuations de température (FT). Les méthodes AD et CT sont en bon accord pour la mesure de la densité de flux de quantité de mouvement. Pour la chaleur sensible, vu sa très grande sensibilité aux erreurs de mesure de température, la méthode AD s'avère inopérante de nuit et fournit des estimations nettement biaisées de jour. La méthode FT fournit uniquement des estimations de la densité de flux de chaleur sensible ; l'accord avec la méthode CT est très bon durant la journée, il est moins bon de nuit, la technique FT ne permettant pas de discerner les flux positifs des flux négatifs. Des trois méthodes, la méthode CT est la seule qui permette une mesure en continu à la fois des densités de flux de chaleur sensible et de quantité de mouvement. L'algorithme de traitement de données associé à cette méthode est décrit en détail dans l'article.
Mots-clés. Covariance de turbulence, fluctuation de température, méthode aérodynamique, mesure de flux.

PRESENTATION OF THE METHODS

The momentum and sensible heat fluxes exchanged by a canopy with the atmosphere are essential processes that determine the physical conditions in which the plants are growing. The first characterizes the action of the wind, the second is an important term of the energy balance of the vegetation. The knowledge of these two variables is therefore essential to describe the microclimate in the canopy. In addition all the exchanges (in particular the transpiration and the photosynthesis) apply following similar processes. The presented methods may therefore, in some extent, be applied to all flux measurements.

Different methods were developed to measure these exchanges. In this paper we will compare three of them: the direct (eddy covariance) method, the aerodynamic method, that is based on the assumption of a flux-gradient relationship and uses mean profile measurements, and the temperature fluctuation method derived from similarity relations.

In a previous paper (Aubinet, 1993), we showed that, in homogeneous conditions, the flux densities are proportional to the covariance of the vertical component of the velocity w [$m \cdot s^{-1}$] and the concentration of the exchanged tracer. In the case of the momentum flux density τ [Pa], the tracer is the horizontal component of the velocity u [$m \cdot s^{-1}$]. In the case of the sensible heat flux H [$W \cdot m^{-2}$] it is the air temperature T [K]:

$$H = \rho_a C_a \overline{w'T'} \tag{1a}$$

$$\tau = -\rho_a \overline{w'u'} \tag{1b}$$

where: ρ_a is the air density [$kg \cdot m^{-3}$], C_a is the air specific heat [$J \cdot kg^{-1} \cdot K^{-1}$], the overbars refer to a time average and the primes to the fluctuation around the average. In the eddy covariance method the flux densities are deduced from the direct measurement of the fluctuation products $\overline{w'u'}$ and $\overline{w'T'}$, according to (1). This requires sophisticated measurement and data treatment systems and became feasible only recently thanks to the development of sonic anemometers and performant computers.

Historically, the aerodynamic method was probably the first method to be used to measure fluxes because it does not require sophisticated material. It is based on a flux-gradient relationship assumption and needs the measurements of mean temperatures and velocities at two heights. The equipment is constituted by two thermometers and two cup anemometers.

The temperature fluctuation method is only valid for the sensible heat measurements. It is derived from the Monin Obukhov similarity theory and has been set up recently to estimate the sensible heat fluxes (Weaver, 1990; Lloyd *et al.*, 1991; de Bruin *et al.*, 1993; Pardo, 1993). The method needs two measurements: the temperature standard deviation and the mean air velocity. The first can be obtained with a simple low inertia thermometer (thermistor, thermocouple or platinum wire), the second with a cup anemometer. We will describe the three methods and present a comparison between them.

THEORY

The Monin Obukhov similarity theory describes the atmospheric processes in the surface boundary layer (*i.e.* the layer situated at the interface between soil and atmosphere where the fluxes can be supposed constant along the vertical). It is widely used in the theoretical presentation that follows. It has been extensively presented in several treatises (Businger, 1973; Panovsky, Dutton, 1984; Arya, 1988). We briefly recall its main features.

This theory assumes that the momentum and sensible heat fluxes in the surface boundary layer are controlled only by the following parameters: the altitude z [m], the momentum flux density τ [Pa], the sensible heat flux density H [$W \cdot m^{-2}$] and the buoyancy g/T, where g is the gravity acceleration [$m \cdot s^{-2}$]. Therefore, by application of the Buckingham similarity theorem, it can be shown that any variable X characterizing these processes can be described as

$$\frac{X}{x*} = f\left(\frac{z}{L}\right) \qquad (2)$$

where z/L, known as the stability parameter (L is the Obukhov length), writes

$$\frac{z}{L} = \frac{-H}{\rho_a C_a} \frac{kgz}{T} \frac{1}{(\tau)^{3/2}} \qquad (3)$$

where k is the von Karman constant (≈ 0.4). The scaling factor $x*$ in (2) has the same dimension as the variable X. In particular, the following scaling factors are currently used

$$T* = \frac{1}{u*} \frac{-H}{\rho_a C_a} \qquad (4a)$$

in the case of temperature, and

$$u* = \sqrt{\frac{\tau}{\rho_a}} \qquad (4b)$$

that is known as the friction velocity.

The stability factor has the opposite sign than the sensible heat flux and characterizes the thermal stratification of the boundary layer: it is positive (negative) when the air temperature increases (decreases) with height. The boundary layer is then said to be "(un)stable". It reaches zero when the temperature gradient decays: the boundary layer is then said to be "neutral". In unstable conditions, the stability parameter has been shown to be equal to the Richardson number, defined as

$$Ri = \frac{g}{T} \sqrt{z_1 z_2} \ln\left(\frac{z_1}{z_2}\right) \frac{(T_2 - T_1)}{(u_2 - u_1)^2} \qquad (5)$$

where $T_{1(2)}$ and $u_{1(2)}$ are respectively the air temperature [°C] and the wind speed [$m \cdot s^{-1}$] at the height $z_{1(2)}$. This shows that estimations of the boundary layer stability are possible by using measurements of mean air temperature and mean wind speed.

The eddy covariance method

The eddy covariance (EC) method consists in directly measuring the covariance between the vertical component of the velocity and the tracer concentrations and deducing the flux densities using the relation (1). It seems to be the simplest method but its technical requirements are very high: it needs high frequency (about 20 Hz) measurements of the air velocity and temperature; it also needs a three directional measurement of velocity. We will detail materials and software features in **Eddy covariance measurements** paragraph.

Note in addition that this method, like the two others, is submitted to the hypothesis of horizontal homogeneity: its validity is limited to homogeneous crops, in particular the measurement system must be placed at a sufficient distance of the edges. The currently recommended upwind fetch is of the order of 100 times the height of the measurement point above the effective surface.

The aerodynamic method

The aerodynamic (AD) method supposes that the tracer flux is proportional to the mean tracer concentration and mean velocity gradient. Using this hypothesis, that is valid only above the crop, and applying the Monin Obukhov similarity theory, it is possible to obtain a relation between the flux densities and the mean air temperature and velocity differences between two heights (Arya, 1988; Businger, 1973):

$$H\mathrm{AD} = -\rho_a\, C_a \frac{k^2 (T_2 - T_1)(u_2 - u_1)\, F_t\, F_u}{\ln (z_2/z_1)^2} \qquad (6a)$$

and

$$\tau\mathrm{AD} = \rho_a \frac{k^2 (u_2 - u_1)^2\, F^2_u}{\ln (z_2/z_1)^2} \qquad (6b)$$

where F_u and F_t are non-dimensional similarity functions of the stability parameter. Their values approach 1 in neutral conditions. Several formulations of these functions were proposed. All were based on experimental measurements and do not differ significantly from one another. We choose the formulation proposed by Arya (1988) valid under unstable conditions:

$$F_t = F_u^2 = (1 - 15\ Ri)^{1/2} = (1 - 15\ z/L)^{1/2} \qquad (7)$$

From relations (5–7), it appears that the momentum and heat flux densities can be deduced from measurements, at two heights, of the mean air temperature and wind velocity. It is an advantage of the AD method to require simple material, (only mean values are required). However, it is based on the flux gradient relationship and is only valid in the boundary layer that develops above the canopy. We discussed (Aubinet, 1993) the failures of the method inside canopies. In addition above high canopies (in forest, for example), the velocity and tracer concentrations gradients are very low. Reliable estimations of flux densities can then be obtained only for important height differences between measurement points. Thus AD method requires very high masts and fetch requirements will be fulfilled only over very wide canopies.

The temperature fluctuation method

Applying the Monin Obukhov similarity theory to the temperature standard deviation σ_T [K], we find

$$\frac{\sigma_T}{|T*|} = f\left(\frac{z}{L}\right) \qquad (8)$$

In unstable conditions, when fluxes are controlled by both free and forced convection, the shape of $f(z/L)$ cannot be predicted by the theory, but in high unstability ($z/L < -1$) when convection is entirely free, we can write

$$\frac{\sigma_T}{|T*|} = f\left(\frac{z}{L}\right) = C_1\left(-C_2\frac{z}{L}\right)^{-1/3} \qquad (9)$$

where C_1 and C_2 are constants. The particular writing of (9) will be justified by the mathematical treatment we will operate later. The values of C_1 and C_2 are deduced from experimental measurements (Businger, 1973; Panofsky, Dutton, 1984; Arya, 1988). In this work we will use the same values as de Bruin *et al.* (1993) ($C_1 = 2.9$ and $C_2 = 28.4$).

On the other hand, when the boundary layer approaches neutral conditions (exchanges controlled only by forced convection), the temperature standard deviation does not depend on the stability parameter, and we have

$$\frac{\sigma_T}{|T*|} = f\left(\frac{z}{L}\right) = C_1 \qquad (10)$$

In introducing the expression (3) of z/L in (9), we find, after rearrangement

$$\frac{H}{\rho_a\, C_a} = \left[\left(\frac{\sigma_T}{C_1}\right)^3 \frac{\mathrm{kg}z}{T}\, C_2\right]^{1/2} \qquad (11a)$$

which could also write

$$\frac{H}{\rho_a\, C_a} = \frac{\sigma_T}{C_T}\, u_f* \qquad (11b)$$

where the velocity u_f* is defined as

$$u_f* = \left[\left(\frac{\sigma_T}{C_1}\right) \frac{\mathrm{kg}z}{T}\, C_2\right]^{1/2} \qquad (12)$$

The relation (11a) shows that, in natural convection conditions, the sensible heat flux density can be immediately deduced from measurements of the temperature standard deviation provided the height and the absolute temperature are known. It is the base of the temperature fluctuation (TF) method. It was used successfully by Lloyd *et al.* (1991) to estimate heat fluxes above bare soils, millet crop, fallow savannah and tiger bush in Niger where the free convection approximation is valid. Nevertheless it should not be the case in temperate areas where the turbulence is often controlled by both forced and free convection.

However, in neutral conditions, a relation similar to (11b) can be found in combining (4a) and (10):

$$\frac{H}{\rho_a C_a} = \frac{\sigma_T}{C_1} u_n* \qquad (13)$$

where u_n* is the friction velocity in neutral conditions.

Noting that (12) and (13) have a similar shape, the sole difference between them being the definition of the friction velocity, de Bruin *et al.* (1993) proposed a general relation, valid for all the instable and near neutral conditions, with have the same shape but where the velocity should be estimated by interpolation between the natural convection value and the neutral value. They proposed the following interpolation formula

$$\frac{H}{\rho_a C_a} = \frac{\sigma_T}{C_1} \left(u_n^{1/p}* + u_f^{1/p}*\right)^p \qquad (14)$$

In a further research, de Bruin (1994) gives more precise description of p and demonstrate that its value depends on the height of measurement and on the roughness length. In our case the value of p is 0.42.

The last problem is to characterize the value of u_n*. It is the friction velocity in neutral conditions and can be deduced from one velocity measurement at a given height. Indeed, in neutral conditions, the Monin Obukhov similarity theory shows that (Businger, 1973; Panofsky, Dutton, 1984; Arya, 1988)

$$u_n* = \frac{k\ u(z)}{\ln(z/z_0)} \qquad (15)$$

where z_0 is the roughness length [m].

The temperature fluctuation method is based on relations (12), (14) and (15). It makes possible to estimate sensible heat flux densities from measures of the temperature standard deviation and of the mean velocity provided the mean temperature and the roughness length of the canopy are known. One measurement point is sufficient (like for EC method) which makes the measurement possible above tall canopies. However the domain of validity of the method is the same as for the aerodynamic method: it is valid where the similarity theory applies *i.e.* in the boundary layer above the canopy. In particular, it fails inside the canopy. Moreover it is rigorously valid only in unstable conditions. Let us note finally that the fetch requirements are the same as for the other methods.

MATERIAL AND METHODS

Site characteristics

Measurements were performed during spring 1994 above a fallow terrain covered by grass. The terrain had a slope of 5% in the NW direction. The instruments were placed at least at 150 m of the edge of the terrain in the dominant upwind direction (W). The crop height was about 0.3 m.

Eddy covariance measurements

In our system, the vertical and horizontal components of the velocity as well as the air temperature were measured with a 3D sonic anemometer (GILL SOLENT 1012-K-055) placed at 2.2 m height. A description of this apparatus was given in a previous paper (Aubinet, 1993). It gives measurements of the three components of the velocity in a coordinate system linked to the apparatus. In addition it measures the speed of sound C [$m \cdot s^{-1}$] from which temperature may be deduced according to (Kaimal, Gaynor, 1991)

$$T = \frac{C^2}{403} \qquad (16)$$

A classical eddy covariance procedure should include a measurement period followed by a computation period during which the covariances are calculated. However the latter period is quite long (after a 15 minutes measurement period, the program has to treat a 18,900 × 4 matrix) which is incompatible with the need for continuous measurements, crucial in micrometeorological studies. On the other hand, the storage of the raw data for further treatment should require too much memory to be achievable with a small computer. We developed a BASIC program that enables to pass round this double obstacle by means of a running mean algorithm. It works according to the following procedure.

– Initialisation of the running mean: during one given period (the time constant of the running mean which order of length is several minutes), temperature and velocity vector are measured at 21 Hz. Their means are computed.

– Measurement and on line summation: every 21th of a second, the instantaneous temperature and velocity components are measured and summed; the temperature and velocity fluctuations in the reference system of the anemometer are computed by substracting the corresponding running mean; the fluctuation products are computed and summed up; the running mean is refreshed with the new measurements.

– Computation and coordinate rotation: after each measurement period, the mean temperature and velocity vector as well as the variances and covariances are immediately given in the reference system by the sums carried out during the measurement period. Their components in a system linked to the mean wind velocity (x parallel to the mean wind velocity, y horizontal and normal to it, z normal to x and y) are computed using a coordinate rotation. The latter can be performed rapidly without interruption of the measurements. In some cases, complete data were recorded for a deeper analysis. Let us now detail some key points.

Running mean. The role of the running mean is not only to facilitate the on line computation of covariance but also to detrend the data: it acts like a high pass filter that suppresses the low frequency contributions (alternance of cloudy and clear passages, increases or decreases of temperature, etc.) that are independant of the exchange process.

The running mean was refreshed according to (McMillen, 1986)

$$X_{new} = X_{old}\left(1 - \frac{\Delta t}{t}\right) + x\,\frac{\Delta t}{t} \qquad (17)$$

where: X_{new} is the new mean, X_{old} is the old mean, x is the instantaneous measurement, Δt is the time interval between two measurements, t is the time constant of the running means.

One problem is the determination of the time constant of the running mean. No clear rule can be given for this choice, time constants of 100 to 200 s are often proposed for micrometeorological purposes. On some days, we compared measurements filtered with two different time constants (60 and 240 s) and found practically no differences between the two results. We use then systematically a 60 s time constant. Let us note that this choice depends on the characteristics of the terrain and must always be adapted (Moncreiff, personal communication).

Change of axis. The more currently used coordinate rotation (McMillen, 1986; Baldocchi *et al.*, 1988) consists in two rotations around the y and z axes that makes the x direction coincide with the mean velocity vector. The elements of the rotation matrix are deduced from the mean velocity components. Let us note however that, rigorously, three rotations should be necessary: a x rotation should be added so that the lateral momentum covariance $\overline{w'v'}$ decays. This third rotation is more difficult to apply because the rotation angle must be found by trials and errors.

During our measurements we applied systematically the two coordinate rotation changes and found, on some windy days, very high values of $\overline{w'v'}$ (sometimes greater than $\overline{w'u'}$). We applied then a third rotation and found that very low angles were sufficient to nullify $\overline{w'v'}$. This shows that the lateral momentum covariance is very sensible to a verticality error. Nevertheless the other covariances (including $\overline{w'u'}$) were practically insensitive to the third rotation as shown on **figure 1**. In consequence we systematically used the simple two rotations coordinate change which seems sufficient.

Corrections on sonic temperature. The velocity of sound measurement, used for the temperature estimation according to (16), can be perturbated by humidity fluctuations or by lateral wind puffes. Corrections to remove these perturbations are applied. For the humidity, we use the correction proposed by Laubach *et al.* (1994) which does not differ significantly from those proposed by Kaimal and Gaynor (1991) or Busch (1973):

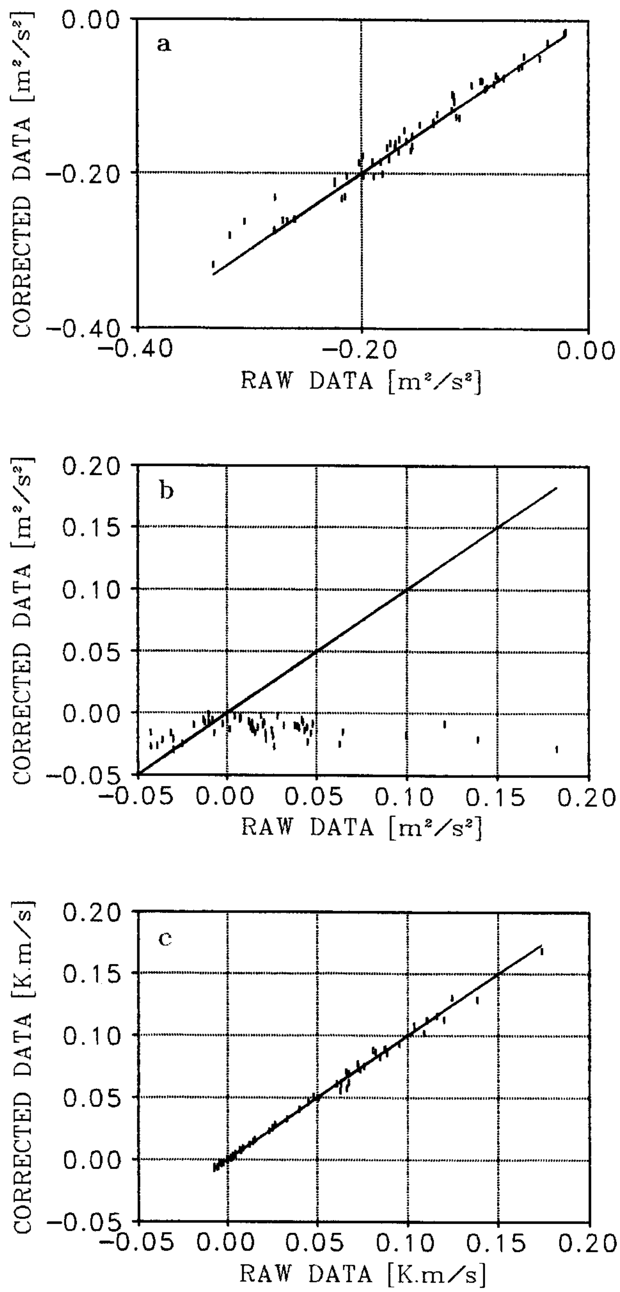

Figure 1. Impact of the verticality error (neglecting the x axis rotation) on the flux density estimations. Dots: corrected data (with third rotation); line: raw data (without third rotation). **(a)** Vertical momentum flux density ($\overline{u'w'}$). **(b)** Lateral momentum flux density ($\overline{v'w'}$). **(c)** Heat flux density ($\overline{w'T'}$).

$$H_{cor} = H_{unc}\,(1 - 0.06/\beta) \qquad (18)$$

where β is the Bowen ratio [–] defined as the ratio of sensible to latent heat flux densities and the indices refer to the corrected or the uncorrected flux densities.

For the lateral wind component, we used (Kaimal, Gaynor, 1991)

$$H_{cor} = H_{unc} + \rho_a C_a \frac{2\,\bar{u}\,\overline{u'w'}}{403} \qquad (19)$$

In our experiment, the first correction leads to modifications of the sensible heat up to 10%; the second was significative only during windy periods where it reached 5% (**Figure 2, a** and **b**).

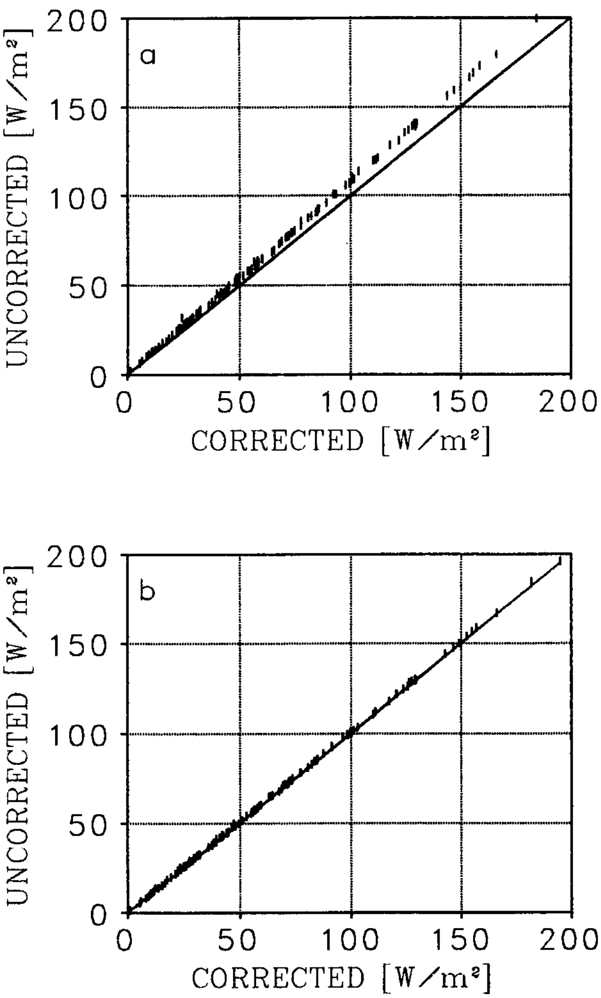

Figure 2. Impact on the EC estimation of heat flux density of the errors on sound velocity. Dots: corrected data; line: raw data. **(a)** Humidity error. **(b)** Lateral flux error.

Filtering. One key point when recording turbulence is to be sure that the measurement frequency is sufficient to detect all the meteorological processes at work in the boundary layer. Indeed the measurement system acts as a filter that erases the processes occurring at too high or too low frequencies. A method to determine the frequency range where the processes take place is to compare its cospectrum to the bandwidth of the measurement system.

The errors introduced in eddy covariance measurements by filtering were detailed very completely by Moore (1986). We apply his analysis on our system to compute its bandwidth which is given on **figure 3**. It shows that the system records all the processes taking place between 10^{-2} and 2 Hz practically without attenuation. Beyond these limits, the sensitivity of the system decays rapidly. The more restrictive limitations are due at low frequency to the running mean, and at high frequency to the spatial averaging (losses of high frequency variations because the measurement is performed over a finite path length). We showed earlier (by changing the time constant of the running mean) that the cut low frequency had no impact on the measurements. On the other hand the impact of the cut high frequency is weak. Indeed the cospectra of $\overline{w'u'}$ and $\overline{w'T'}$ decay at frequencies that are markedly lower than the cutoff frequency of the system. Typical cospectra of $\overline{w'u'}$ and $\overline{w'T'}$ are shown on **figure 3**. However these results cannot be extended to experiments made under other conditions: the frequency range where the processes take place varies with the roughness of the crop and the measurement height.

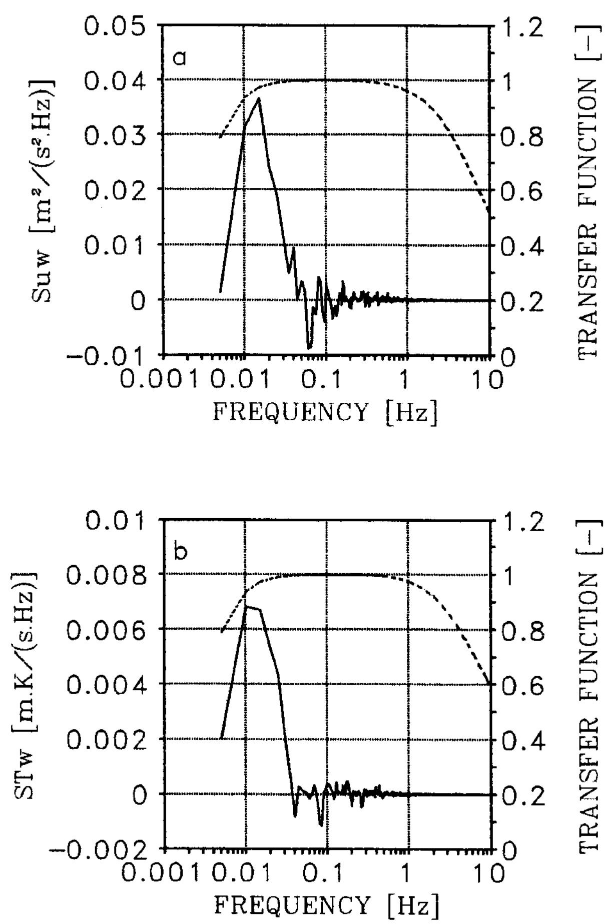

Figure 3. Transfer function (dashed line) of the EC system compared to: **(a)** the momentum cospectrum (Suw) (solid line), **(b)** the heat cospectrum (STw) (solid line).

Aerodynamic measurements

Vertical profiles of velocity were measured with cup anemometers placed at two heights (0.8 and 8.5 m). Before measurements they were calibrated by comparison with the sonic anemometer in a wind tunnel and in open air.

Air temperatures and humidities were measured by thermopsychrometers made of PT1000 probes placed in tubes with double radiation shading. The ventilation was performed by a fan in the internal tube, it was natural in the external enclosure. The probes were calibrated in laboratory (calibration bath) by comparison with a reference (Thermoelectric MRT-43602-Pt100-250-1500-classA). The error in laboratory was lower than 0.05°C. However in the field, in spite of the double shading, it can reach 0.2°C under strong solar radiation.

The mean values of velocities and temperatures were computed and stored on EPROM. Every 15 min, in synchronism with the EC measurement, their values were picked up and stored on a personal computer.

We observed that the AD method is very sensitive to some measurement errors. Indeed, differences between measurements, on which the method is based, have often the same order of magnitude than the error that affects the individual measurements. It is especially true when the gradients are low. Consequently the two measurement heights must be chosen so that the temperature and velocity differences are as big as possible. A compromise must be found between these requirements and fetch limitations.

Velocity errors. Velocity errors can be due to onset problems (at too low velocity, the cup anemometers do not start), to wake errors (for certain wind directions, the anemometers can be in the wake of the supporting mast) or to inertia errors (the cups keep on to turn after short puffes). During these periods flux measurements are not available or biased.

In addition, the wind velocities can be affected by random errors. Their amplitude is of about 0.1 $m \cdot s^{-1}$. The impact of these errors on flux density errors is amplified by a factor that depends on the stability. The relative error on velocities is multiplied by a factor 1 (neutral) to 2 (natural convection) for the heat flux. It is multiplied by a factor 2 (neutral) to 2.8 (natural convection) for the momentum flux.

Temperature errors. The estimation of the temperature difference is particularly tricky. It is of the order of some tenth of a degree (the maximum recorded value by a very sunny day was 3°C) where the error can reach, as said before, 0.2°C in spite of a lot of experimental precautions. This means that the relative error on the temperature difference is currently of the order of several tens of percents. From our point of vue, this is the essential failure of the aerodynamic method.

Temperature fluctuations measurements

The temperature fluctuations were measured by a low inertia thermistor (SIEMENS K 19) put in a Wheatstone bridge in order to linearize the response of the probe. The output voltage of the bridge was amplified in order to provide 4–20 mA output. The system was calibrated by comparison with the reference PT100 probe in the laboratory. The high frequency acquisition was performed through the sonic anemometer. The standard deviation computation was realised on line by means of a running mean algorithm.

The air velocity was measured with a cup anemometer. The two apparatus were placed at 2 m height.

RESULTS

Comparison between EC and AD methods

Heat and momentum flux densities were measured between 25 May and 1 June simultaneously with EC and AD method. We removed from measurements those corresponding to stable periods ($z/L > 0$) or which were affected by onset or wake errors. The evolutions of the heat and momentum flux densities measured by both methods during two selected days are represented on **figure 4**.

The agreement is quite satisfactory for the momentum flux densities. For the sensible heat, the parallelism between the two estimations is obvious, the two methods accounting simultaneously for the same sharp variations, but the AD method overestimates clearly the flux densities. This is especially manifest on **figure 5** where we present a direct comparison between EC and AD estimations for both flux densities.

The discrepancy is due to an error in temperature measurement that affects the AD method. Two arguments lead to this conclusion. First the heat flux predictions by the AD method are unrealistic by clear days: in May–June, the global solar incident radiation reaches 1,000 $W \cdot m^{-2}$, of which about 25% are reflected by the soil. The net far infrared radiation loss is about 100 $W \cdot m^{-2}$. The order of magnitude of the Bowen ratio that we measured was 0.2 to 0.5. In these conditions the sensible heat flux density should not exceed 220 $W \cdot m^{-2}$. On **figure 4** we see that the AD estimations of the sensible heat flux density largely exceed this value. Secondly, a better agreement between the two methods is obtained when replacing in the AD method the temperature measurement at 0.8 m by another one made at 2 m height with a third probe. This suggests clearly that the 0.8 m probe was affected by a systematic error. Estimations of the heat fluxes on the basis of measurements made at 2 and 8.5 m are however very imprecise, the temperature difference between these two heights being too small.

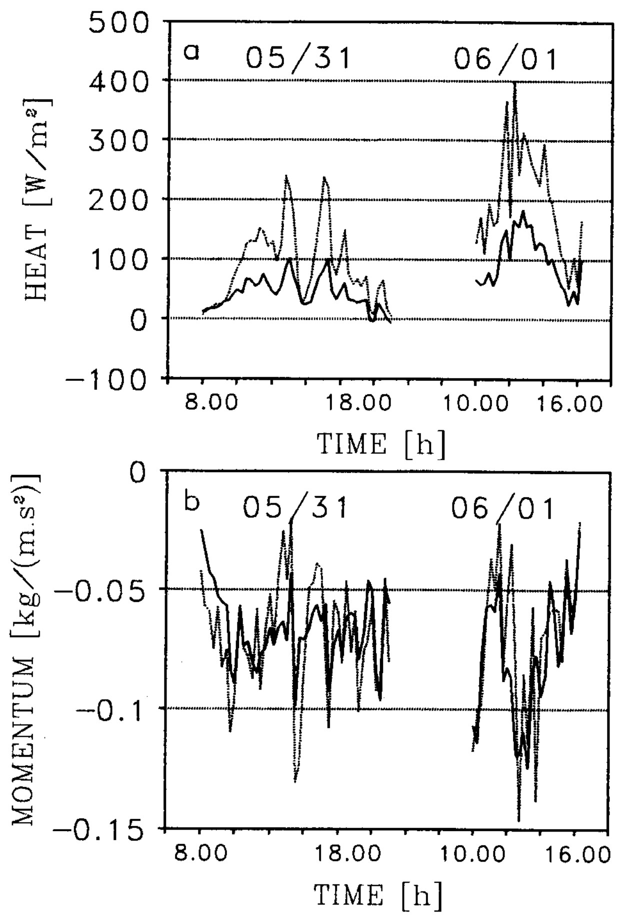

Figure 4. Evolution of heat **(a)** and momentum flux densities **(b)** during selected periods (30 and 31 May, 1 June). Solid line: AD estimations; dotted line: EC estimations.

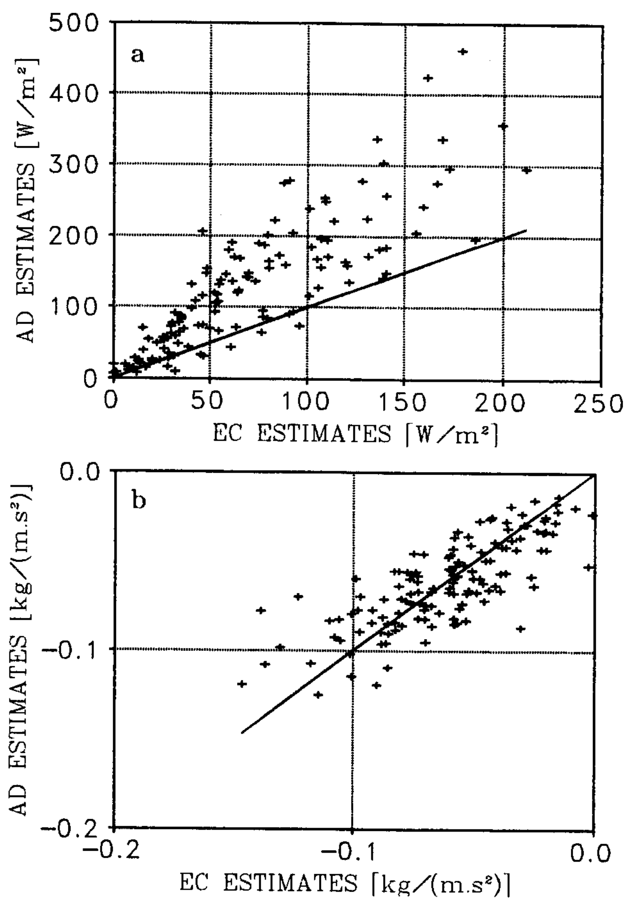

Figure 5. Comparison between AD and EC estimates of heat **(a)** and momentum flux densities **(b)** between 25 May and 1 June.

Let us recall that we took care in reducing the temperature error and that in fact this was small (some tenth of a degree on the 0.8 m probe). In particular, such an error cannot be detected by a simple examination of the temperature profiles. However under low temperature gradient it is sufficient to induce errors up to 100% when using the AD method. As it is impossible to completely remove systematic errors on the probes, an interesting alternative should be to swap the probes during the measurement. A system based on this principle has been set up and used successfully during the 1995 campaign.

Comparison between EC and TF methods

EC and TF estimations of sensible heat flux were measured simultaneously between 17 and 24 June. The measurements were performed continuously, night and day, during four days with an interruption due to a system defect. The comparison between the two methods is given at **figure 6**. The agreement is particularly good during all instable periods. The slope of the regression of TF vs EC estimations is 0.927 ($Rsq = 0.884$).

However TF method gives only absolute values of the heat flux (rigorously, it can be used only under neutral and instable conditions) which explains that during night periods discrepancies appear between the two estimations. It is nevertheless interesting to note that at these moments the TF estimations are practically the mirror image of the EC estimations. This suggests that the TF method could be extended to stable conditions even if it was not its aim previously.

CONCLUSIONS

Three methods for estimating sensible heat and momentum flux densities were tested over a fallow crop.

The aerodynamic method gave quite poor results: it is not only unfeasible by low winds (particularly during stable periods) but also very sensitive to temperature errors. The system could markedly be improved in eliminating the systematic errors. One promising way is to swap the position of the thermometers during every measurement. Such a system has been set up now.

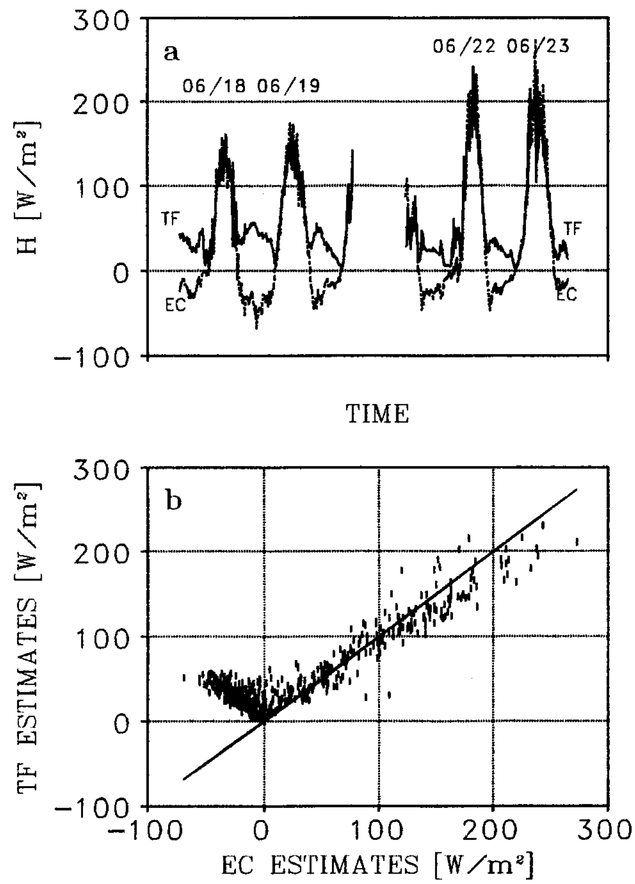

Figure 6. Comparison between TF and EC estimates of heat flux density between 17 and 24 June.
(**a**) Time evolution. (**b**) Direct comparison.

The temperature fluctuation method gave satisfactory results not only during unstable but also during stable periods. The greatest asset of the method is the simplicity of the probes and their low price. However it is only limited to heat measurements in the surface boundary layer where similarity theory applies. Moreover, a shortcoming of the system is the great brittleness of the temperature probe.

The eddy covariance method gave the best results. It can be used in all circumstances provided that the fetch conditions are fulfilled. In addition, the system operated during all the installation and measuring periods (several months) without any failure. This system is therefore the best that can be found for long term measurements. Its main drawback is its prohibitive price.

All three methods can be extended to the measurement of other fluxes (water vapour, carbon dioxide) provided that the system should be completed by an appropriate concentration analyzer. In the case of the aerodynamic method, it must be of high precision to detect very low concentration gradients. This makes the method still more tricky for the measure of CO_2 fluxes than for the sensible heat fluxes. In the case of eddy covariance and fluctuation methods, the apparatus must have fast response.

An eddy covariance system using together a CO_2/H_2O analyzer and a sonic anemometer has been set up and put into service in 1996 by the Department of Physics, of the University Faculty of Agricultural Sciences, Gembloux. This new measurement campaign fits into the frame of an European network for the measurement of long term CO_2/H_2O fluxes of forests (EUROFLUX project).

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