Current Micrometeorological Flux Methodologies with Applications in Agriculture

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INTRODUCTION

History has shown that many of the technological advances in micrometeorological measurement methods and techniques were facilitated by agronomic research into understanding plant-water relationships and photosynthesis (Deacon and Swinbank, 1958; Lemon, 1960; Suomi and Tanner, 1960; Tanner, 1960). Even today, many research programs are focusing on a better understanding of the water balance on regional and continental scales with a goal of providing useful probabilistic forecasts that will lead to more effective planning of the utilization of water resources in areas where water is limited or stricken by drought (Lawford, 1999). The need to understand and predict components of the water and carbon cycle on seasonal and annual time scales has pushed the micrometeorological technology to work not only for short duration intensive experiments, but also for much longer periods (Baldocchi et al., 2002; Running et al., 1999; Grelle and Lindroth, 1996).

Many agricultural research activities include short-term intensive experiments in which the emission or deposition of some chemical species is to be quantified. Along with the more common species like water vapor and CO_2, fluxes of other chemical species have become important in agriculture. They include nitrogen compounds such as NH_3, HNO_3, NO, other greenhouse gases such as CH_4 and N_2O, pollutants such as O_3 and SO_2, and many pesticides (Majewski, 1989). Measured fluxes are also being integrated over the year to compile annual budgets of water and carbon. Potential biases in the measured fluxes that could lead to potentially large errors in these annual budgets has spurred several efforts to re-examine the methodologies of current flux measurement techniques (Massman and Lee, 2002), and how the data should be processed to insure the highest data integrity (Foken and Wichura, 1995). Although key findings and recommendations from these papers will be discussed here, reading these papers is encouraged.

Here, we will describe the various micrometeorological methodologies that are *currently* used to determine the vertical turbulent fluxes of water, CO_2, and other scalar entities on both short and long time scales, incorporating the latest strategies on flux measurement techniques. There are many well-written historical papers and book chapters on micrometeorological methodologies (Webb, 1966; Kanemasu et al., 1979; Hicks, 1984). The reader is encouraged to read these references as these

provide basic information on the fundamental processes of air-surface exchange, which is necessary to appreciate and understand the application of the various measurement methodologies.

For many agricultural applications, micrometeorological methods are preferred since they are generally non-intrusive, can be applied on a semi-continuous basis, and provide information about vertical fluxes that are aerially averaged on scales ranging from tens of meters to several kilometers, depending on the roughness of the surface, the height of the instrumentation, and the stability of the atmosphere surface layer. Interpretation of the data obtained from these methods are also constrained by the assumptions of stationary in the scalar concentration field, horizontal homogeneity of the surface conditions, and relatively flat terrain. These constraints insure that the vertical turbulent flux is due to the scalar source/sink at the surface and not by advective horizontal mean or turbulent fluxes.

The most commonly used micrometeorological methods can be separated into four categories. They are:

1. eddy covariance
2. flux-gradient
3. accumulation
4. mass balance

Each of these techniques is suited for applications that depend on the scalar of interest and surface type, over which the flux is to be measured. For example, the eddy covariance method can only be used for trace species for which fast response instrumentation is available. Likewise, flux gradient or accumulation methods are often used in systems for which the atmospheric concentrations can be determined with a high degree of accuracy using either slow response sensors or accumulation devices such as flasks, canisters, filter packs, or annular denuders.

There are other methodologies that have been used to measure fluxes including the variance method (Padro, 1993) and the inertial dissipation technique (Fairall and Larsen, 1986). These methods are usually applied in very special circumstances and cannot be necessarily generalized for all trace gases and environmental conditions. In this chapter, we will discuss the basics and strengths of the four methodologies above. Our intent is not to elaborate in detail on the theoretical derivations and complete historical background of these methods, but to discuss and describe how these can be utilized to measure fluxes on time scales that are pertinent to research in agriculture. There are many excellent reviews, both current and historical, that provide excellent background material that need not be repeated here.

2 FLUX MEASUREMENT METHODOLIGIES

All flux measurement strategies begin with examining the conservation equation for a scalar *s* that is usually applied to a control volume. For the discussion here, the volume will encompass the region over and including the surface (crop, grass, forest, etc.) of interest (Figure 1).

$$\frac{\partial s}{\partial t}+\frac{\partial us}{\partial x}+\frac{\partial vs}{\partial y}+\frac{\partial ws}{\partial z}=S(x,y,z) \tag{1}$$

The instantaneous fluxes in and out of the volume are comprised of the streamwise (*x*), crosswind (*y*), and vertical (*z)* directions, denoted by *us, vs,* and *ws*, respectively, where *u, v,* and *w* are the corresponding wind velocity vectors, and *S(x,y,x)* is the source/sink term. Applying Reynolds decomposition (i.e. $s=\bar{s}+s', w=\bar{w}+w'$, etc.),

averaging, and incorporating the continuity equation,

$$\bar{s}\left(\frac{\partial \bar{u}}{\partial x}+\frac{\partial \bar{v}}{\partial y}+\frac{\partial \bar{w}}{\partial z}\right)=0 \tag{2}$$

the expression becomes,

$$\frac{\partial \bar{s}}{\partial t}+\bar{u}\frac{\partial \bar{s}}{\partial x}+\bar{v}\frac{\partial \bar{s}}{\partial y}+\bar{w}\frac{\partial \bar{s}}{\partial z}+\frac{\partial \overline{u's'}}{\partial x}+\frac{\partial øverline v's'}{\partial y}+\frac{\partial \overline{w's'}}{\partial z}=\bar{S}(x,y,z) \tag{3}$$

where the second, third, and fourth terms are the mean flow advective components and the last three terms are divergences or gradients in the turbulent fluxes. For conditions that are near steady state ($\partial \bar{s}/\partial t=0$), and horizontally homogeneous, the steamwise and crosswind gradients are assumed equal to 0. With no net horizontal fluxes (mean or turbulent) into the volume, we have

$$\bar{w}\frac{\partial \bar{s}}{\partial z}+\frac{\partial \overline{w's'}}{\partial z}=\bar{S}(z) \tag{4}$$

where the source term ($\bar{S}(x,y,z)$) represents the changes to the vertical turbulent flux that result from the elements from within the canopy. Over flat terrain, it is usually assumed that there is no net vertical velocity (i.e. $\bar{w}=0$), and the vertical advection term drops out. For the discussion here, similar assumptions will apply. However, later in the eddy variance section, the assumptions of a non-zero vertical velocity, especially as it pertains to coordinate rotation and mean removal, will be revisited. Integrating from the ground to a measurement point at a height *z* above the canopy, the vertical turbulent flux (F_s) above the surface or canopy is

$$F_s=\overline{w's'}\text{\textbf{\textbf{\textbf

1. Eddy Covariance

The eddy covariance method is, in general, the most preferred because it provides a direct measure of the vertical turbulent flux (F) across the mean horizontal streamlines (Swinbank, 1951) providing fast response sensors (≈ 10 Hz) are available for the wind vector and the scalar entity of interest. The flux is usually described simply by

$$F_s = \overline{w's'} \tag{6}$$

where w is the vertical velocity and s is the concentration of the scalar of interest. The primes in Eq. 6 denote a departure from a temporal average (i.e. $w' = w - \overline{w}$) and the overbar represents a time average operation. Strictly speaking, the vertical turbulent flux is the covariance of a time series of the vertical wind velocity and concentration of a scalar entity over some time interval. In practice, the averaging period should be long enough to capture all of the eddy motions that contribute to the flux, which has traditionally been for 30 - 60 minute periods. For aerodynamically rough surfaces in low wind conditions, longer averaging periods may be necessary to capture all frequencies contributing to the flux (Finnegan et al., 2002). If the flux being measured consists of an emission rate of ammonia over a field recently fertilized with granular urea, then the upward moving eddies ($w' > 0$) would contain concentrations of ammonia that were greater than the average ($s' > 0$) and the downward moving air parcels ($w' < 0$) would contain ammonia concentrations that were below average ($s' < 0$), with both generating a positive covariance ($\overline{w's'} > 0$). Raw velocity and trace gas signals from an experiment over soybeans are shown in Figure 2. The coherency in the temperature and water vapor signatures are evident as well as the negative correlation between water and CO_2, especially after 30 s. The lag of the O_3 signal (closed path sensor) was not accounted for, but with the soybeans taking up both CO_2 and O_3, the lag is apparent, especially between 30 and 40 s into the time series.

There are several ways in which the covariance can be computed. Reynolds or block averaging, in which the arithmetic averages ($\overline{w}$ and $\overline{s}$) are computed at the end of an averaging period and w' and s' are computed for each element of the time series can be used to obtain the flux

$$\overline{w's'} = \sum (w - \overline{w})(s - \overline{s}) \tag{7}$$

Computationally, this can be done with a single pass through the data from

$$\overline{w's'} = \frac{1}{n}\sum ws - \frac{1}{n^2}\sum w \sum s \tag{8}$$

where the overbar again denotes a time average operation. For scalar concentrations, block averaging is appropriate if the assumption of stationarity is not violated ($\frac{\partial \overline{s}}{\partial t} \approx 0$). Trends (if they are significant) in the s time series that are not believed to represent contributions to the flux can be removed so that the true turbulent deviation can be computed. This strategy has been appropriate and often necessary when using chemical sensors in eddy covariance applications that may have a significant zero-drift of the output signal on time scales similar to the averaging period. Detrending is usually accomplished using a linear detrending scheme or running mean removal. Linear detrending is another form of high pass filtering in which the linear average of *s* and *w* is represented by a linear regression that has been generated from the time

series coincident with the averaging period. Using this strategy, many researchers do a summation of the covariance "on the fly" using a running mean or digital high pass recursive filter to compute the deviations (Loyd, 1984; McMillen, 1988). After initialization, the real-time running mean (c) at each sample interval is computed as

$$c_i = a c_{i-1} + (1-a) c \tag{9}$$

where c_i is the current running mean, c_{i-1} is the running mean from the previous time step, and c is the current instantaneous value. The constant a is derived from the sampling interval (Δt) and desired time constant t as

$$a = \exp(-\Delta t / t). \tag{10}$$

Although there is no prescribed value for t , time constants between 400 and 1000 s are most often used (Loyd et al., 1984; Shuttleworth et al., 1984). Use of a running mean or alternatively a time centered "boxcar" mean can necessarily remove low-frequency contributions to the flux that may comprise a significant fraction of the covariance. Long period fluctuations (10 - 30 min) have been shown to have an important contribution to the covariance for rough surfaces in light wind conditions during the day (Sakai et al., 2001).

Additionally, after the raw variances and covariance's are computed, they are often "mapped" from the coordinate system from which they were measured (the sonic anemometer frame of reference) to coordinate system in which the z-axis is perpendicular to the mean horizontal streamline for the averaging period (Tanner and Thurtell, 1969; Baldocchi et al., 1988). The mapping is performed with the constraints that $\overline{w} = \overline{v} = 0$ in the new coordinate system. In this two-angle rotation, the coordinate system is first rotated about the z-axis till $\overline{v} = 0$. The next rotation is about the new y axis till $\overline{w} = 0$. Using these constaints, the mapping from the raw to rotated wind velocity vectors is

$$u = u_{raw} \cos q \cos h + v_{raw} \cos q \sin h + w_{raw} \sin q \tag{11}$$

$$v = v_{raw} \cos h - u_{raw} \sin h \tag{12}$$

$$w = w_{raw} \cos q - u_{raw} \sin q \cos h - v_{raw} \sin q \sin h \tag{13}$$

where u can be either $\overline{u}$ or u' and the same for v and w, with

$$\sin q = \overline{w}_{raw} / \sqrt{\overline{u}_{raw}^2 + \overline{v}_{raw}^2 + \overline{w}_{raw}^2} \tag{14}$$

$$\cos q = \sqrt{\overline{u}_{raw}^2 + \overline{v}_{raw}^2} / \sqrt{\overline{u}_{raw}^2 + \overline{v}_{raw}^2 + \overline{w}_{raw}^2} \tag{15}$$

$$\cos h = \overline{u}_{raw} / \sqrt{\overline{u}_{raw}^2 + \overline{v}_{raw}^2} \tag{16}$$

$$\sin h = \overline{v}_{raw} / \sqrt{\overline{u}_{raw}^2 + \overline{v}_{raw}^2} \tag{17}$$

By multiplying the correct equations and performing a time average operation, the time average rotated variances and covariances can be obtained. For example, the kinematic momentum flux, $\overline{u'w'}$, is obtained by multiplying Equations 11 and 13,

then applying a time average operator. This gives

$$\overline{u'w'} = \overline{u'w'}_{raw}\cos h(\cos^2 q - \sin^2 q) + \overline{w'v'}_{raw}\sin h(\cos^2 q - \sin^2 q)$$
$$+\overline{w'^2}_{raw}\sin q\,\cos q - \overline{u'^2}_{raw}\sin q\,\cos q\,\cos^2 h$$
$$-\overline{v'^2}_{raw}\sin q\,\cos q\,\sin^2 h - 2\overline{u'v'}_{raw}\sin q\,\cos q\,\sin h\cos h \qquad (18)$$

Similarly, the rotated vertical turbulent flux of a scalar is computed as

$$\overline{w's'} = \overline{w's'}_{raw}\cos q - \overline{u's'}_{raw}\sin q\,\cos h - \overline{v'w'}_{raw}\sin q\,\sin h \qquad (19)$$

This coordinate rotation has traditionally been performed on data at the end of each averaging period, assuming that the flows for each period were terrain following. Applications of eddy covariance methods to non-ideal terrain, in which non-zero mean vertical velocities are often observed, have lead to a re-examination of the procedures for computing variances and covariances (Paw U et a., 2001; Wilczak et al., 2001). They suggest that the fluxes which have been derived from the coordinate system of the sonic should be mapped into an orthogonal system in which the z-axis is perpendicular to the plane of the *mean* streamline, which has been determined over a long time period using many hours of data from all wind directions. This "planar" fit is obtained by plotting $\overline{w}_{raw}/\sqrt{\overline{u}^2_{raw} + \overline{v}^2_{raw}}$ for all wind directions to define the plane of the mean streamline and using a sonic in which the z-axis is parallel to the gravity vector (Baldocchi et al., 2000). Coordinate rotations done in this manner are consistent with the idea of determining the eddy flux in the direction of the axis perpendicular to the mean streamline while at the same time preserving information contained in a non-zero $\overline{w}$ that may contain information on horizontal or vertical advection (Finnigan, 1999; Lee, 1998). Admittedly, this has been more of an issue for flux measurement systems over tall forests in non-ideal terrain and is usually not a serious issue for agricultural related experiments.

Although elegantly simple, there are other considerations necessary in order to obtain the correct flux. The use of sonic anemometers for applications of eddy covariance is now commonplace rather than an exeption. Advances in sonic anemometer technology have made the measure of the streamwise (u), crosswind (v), and vertical (w) components of the wind vector a relatively simple task. There are now several commercial manufacturers of sonic anemometers. Most, in addition to providing a measure of the three components of the wind vector and air temperature (actually the measured temperature is nearly equal to the virtual temperature), have an associated analog to digital (A/D) converter for obtaining the time series of additional scalars. This brings up a key issue with regard to computing the covariance. For proper determination of the flux (F), the time series of w and s should be in phase. The measurement of the vertical wind component should be correlated with the concentration of the scalar in the air parcel at the time the wind measurement was made. For example, many trace gas analyzers are closed path systems and there is often a short time delay from when the air enters an intake tube near the sensing volume of the sonic to the trace gas detection chamber. There may be a small period of time (a few hundred milliseconds) devoted to signal processing before the instrument outputs a signal for the sonic A/D. When the time delay is accurately determined (usually to within 0.1 seconds), the covariance is properly computed by lagging w to be in phase with s (Moore, 1986).

Another important conversion that is necessary for open path scalar sensors where the output signal from the trace gas sensor is proportional to a density, g/m^3, was

reported on by Webb et al., 1980. This conversion arises since the gas density measured by both open and closed path sensors is affected by changes in the air temperature and humidity, which will induce fluctuations in the trace gas density. For example, consider the case of Argon gas, which is a constituent of the composition of air that is neither emitted nor taken up by the surface. During the day, a warm upward moving parcel of air (w' is +, Θ' is +; $\overline{w'\Theta'} > 0$), would be associated with an Argon signal in which the deviation of the density is negative suggesting a negative flux of Argon. The conversion from Webb et al. (1980) is mass consistent and corrects for density fluctuations resulting from changes in the air temperature and humidity. The full corrected flux (F_s) can be expressed as

$$F_s = \overline{w'r_s} + \overline{r}_s\left(\frac{m\overline{w'r'_v}}{\overline{r}_a} + (1+ms)\frac{\overline{w'T'}}{\overline{T}}\right) \tag{20}$$

where m is the ratio of molecular mass of dry air to water vapor; s is the ratio of mean water vapor density to dry air density; r_s is the density of the trace gas of interest; $\overline{w'r'_v}$ and $\overline{w'T'}$ are the water vapor and sensible heat fluxes, respectively; $\overline{T}$ is the mean temperature ($^\circ$K), and **[Trial mode]**is the raw flux. For example, assuming typical atmospheric conditions, the CO**[Trial mode]**flux (in **[Trial mode]**mol m**[Trial mode]**s**[Trial mode]**) becomes **[Trial mode]**where the latent (**[Trial mode]**) and sensible (*H*) heat fluxes are in W m**[Trial mode]**. For a sensible heat flux of 300 W m**[Trial mode]** and a laten heat flux of 100 W m**[Trial mode]**, a raw CO**[Trial mode]**flux of -30 **[Trial mode]**mol m**[Trial mode]**s**[Trial mode]** is adjusted to -23.95 **[Trial mode]**mol m**[Trial mode]**s**[Trial mode]**, whereas a raw CO**[Trial mode]**flux of -10 **[Trial mode]**mol m**[Trial mode]**s**[Trial mode]** is adjusted to -3.95 **[Trial mode]**mol m**[Trial mode]**s**[Trial mode]**, a substantially greater fraction of the raw flux. Whenever chemical sensors are used that measure the density of a chemical species, whether using open or closed path sensors, these adjustments need to be considered. Usually for closed path systems, unless the air is dried, only the latent heat flux adjustment is applied.

1. Flux Corrections

In addition to insuring the time series are in phase, other corrections to fluxes need to be considered. If closed path sensors are used to sample a particular trace gas, then corrections may be necessary to account for the damped fluctuations from air sampling tubes (Leuning and King, 1992; Moncrieff, 1990), sensor separation, and sensor path averaging (Moore, 1986). The damping of concentration fluctuations as a result of using closed path sensors causes an underestimate of the flux, as there is a loss of covariance at the high frequencies. When using closed path sensors, correction are made based on the flow rate, length of tubing from the sample inlet near the sonic to the analyzer, and atmospheric stability. Corrections can be made using the theoretical expressions presented by Massman and Lee (2001), or they can be empirically derived using cospectral similarity (Wilson and Meyers, 2001). By comparing the normalized co-spectra of a flux determined with closed path sensors, with co-spectra derived from open path instruments, the magnitude of the correction can easily be seen. It is important to maintain a constant flow rate thru the system in order to apply the corrections in a consistent manner.

2. Flux Uncertainties with Eddy Covariance Sampling

The eddy covariance is inherently an application of statistics and has associated uncertainties associated with the sampling of the atmosphere in order to obtain an estimate of the flux. Biases in the fluxes can result from using sensors in which there are systematic drifts in the calibration of the trace gas sensors or in the measurement of the velocity vector. This, of course, is very sensor dependent and will not be the focus of the discussion here. Assuming that there are no biases in the instrumentation, uncertainties in the computed mean flux arise from mainly two sources. The first arises from that fact that with any micrometeorological method, we are sampling environmental variables that are inherently stochastic in nature. If all instrumentation were operating perfectly, there would still be at least **[Trial mode]** 10% uncertainty or variation in the measurement of any flux for a 30-minute interval due to the stochastic nature of turbulence and the natural geophysical variability of the environment (Wesely and Hart, 1985). Instrumentation, especially those for trace gases, can have small biases and usually have varying degrees of random noise that are part of the overall signal. Random noise that can occur with the signal from the trace gas analyzer can lead to a larger run-to-run variability in the measured fluxes but do not lead to biases in the fluxes since the random noise is not correlated with the vertical velocity. This inherent noise rejection capability of the eddy covariance method was illustrated by Meyers and Baldocchi (1993) when they showed that the co-spectral signature of the vertical flux of SO **[Trial mode]** was similar to that for H**[Trial mode]**O and CO**[Trial mode]** although instrument noise constituted a large fraction of the variance and power spectra for SO **[Trial mode]**. In a more recent analysis of the sampling error in eddy covariance measurements, Finkelstein and Sims (2001) showed that by including a better model of the autocorrelation function, the mean normalized error from five different field studies ranged from 12% for sensible heat to 31% for O**[Trial mode]** flux. These are roughly 20 to 25% higher than errors from previously published methods.

2. Conditional Sampling (Relaxed Eddy Accumulation, REA)

First introduced by Businger and Oncley (1989), the conditional sampling or REA method is a modified version of the eddy accumulation technique (Desjardins, 1977). In true eddy accumulation, fast response sampling (i.e. eddy covariance) is replaced with fast response accumulation in “up” and “down” draft collectors. The accumulation rate is proportional to the vertical wind velocity. True eddy accumulation is simple in theory but difficult in practice (McMillen and Hicks, 1984) as it requires very responsive and fast acting valves. With the REA method, sampling or accumulation is not proportional to the vertical velocity, but samples are accumulated at a constant rate, in either updrafts (**[Trial mode]**) or downdrafts (**[Trial mode]**) such that the flux (*F*) is given by
[Trial mode] where **[Trial mode]** is the standard deviation of the vertical wind velocity, **[Trial mode]** and **[Trial mode]** are the average concentrations in the “updraft” and “downdraft” accumulators, respectively, and **[Trial mode]** is an empirical constant. Numerous comparisons with eddy covariance data have shown that **[Trial mode]** is **[Trial mode]**0.6 (Businger and Oncley, 1989; Baker et al., 1992). Pattey et al. (1993) found excellent agreement between CO **[Trial mode]** fluxes from eddy covariance and those derived from REA. In a more recent analysis using four scalars, Katul et al. (1996) determined that an average **[Trial mode]** reproduced the eddy covariance fluxes for heat, water vapor, carbon dioxide, and

ozone. In some slight modifications of this approach, a “deadband” of **[Trial mode]**, symmetric about its mean (usually zero) is used in which there is neither sampling in the up or down accumulators (Bowling et al., 1999). With this modified approach, the concentration difference in the updrafts and downdrafts are maximized and results in a lower **[Trial mode]**. In another slight modification, Meyers et al. (2002) used a deadband similar to Bowling et al. (1999) but only switched the valves each second when simultaneously sampling for gases and aerosols with a combination denuder/filter pack sampling systems. The proportionality constant **[Trial mode]** was computed by assuming similarity in the transfer process. This was accomplished by “synthetically” sampling or accumulating the air temperature and water vapor signals from the eddy covariance data; then solving explicitly for **[Trial mode]** from using these averages, and the eddy covariance flux and **[Trial mode]** over the sampling interval, which ranged from 2 - 4 hours. Using a deadband of **[Trial mode]**, values of **[Trial mode]** were found which are consistent with the findings of Bowling et al. (1999). The attractiveness of this approach is that it can be applied to chemical species for which fast response instrumentation is not yet available. Sampling can be done in evacuated canisters, filter packs, flasks, or any type of chemical scrubber that can be used to measure the concentration to a high degree of accuracy (Guenther et al., 1996; Bowling et al., 1998).

3. Gradient Methods

There are several gradient methods that are currently used to assess the vertical turbulent flux. We will present and discuss the Bowen-ratio energy balance (BREB), modified Bowen-ratio, and aerodynamic methods. These methods rely on the relationship between turbulent fluxes and vertical time average gradients, much in the same way molecular diffusion can be described by a diffusivity and concentration gradient. These methods differ in how the gradients are used to obtain the flux.

1. Bowen-Ratio Energy Balance (BREB)

The basis for the BREB method is that the local energy balanced is closed in such a way that the available net radiative flux (**[Trial mode]**) is strictly composed of the sensible (**[Trial mode]**), latent (**[Trial mode]**), and ground heat (**[Trial mode]**) flux; **[Trial mode]** Inherent in this assumption is that all other storage terms such as those related to canopy heat storage or photosynthesis are neglible. Next, if **[Trial mode]** can be expressed as the product of a vertical gradient of temperature (**[Trial mode]**) over a given height interval (**[Trial mode]**) and a turbulent eddy diffusivity for heat (**[Trial mode]**)
[Trial mode] and similarly, the evaporation (**[Trial mode]**) can be expressed as the product of a vertical gradient of specific humidity (**[Trial mode]**) over the same height interval as the air temperature and it’s turbulent diffusivity (**[Trial mode]**), then the Bowen ratio (**[Trial mode]**) can be expressed as **[Trial mode]** assuming the turbulent diffusivities for heat and water vapor are equal. **[Trial mode]** and **[Trial mode]** are the specific heat of air at constant pressure and the latent heat of evaporation, respectively, and are used to convert the measured units into an energy flux (W/m**[Trial mode]**) and **[Trial mode]** is the air density. The evaporation rate **[Trial mode]** can then be determined as **[Trial mode].** To apply these measurements to obtain the fluxes of other trace species, the eddy diffusivity can be solved for and applied to measurements of vertical gradients made over the same height interval. This method has been successfully applied to obtain seasonal (Angell et al., 2001) and

annual fluxes of CO **[Trial mode]** over grasslands in the western part of the U.S. (Dugas et al., 1999). The limitations of this and other gradient methods is that they are generally not well suited to aerodynamically rough canopies (i.e. forests) because the vertical gradients are often very small and difficult to resolve. Small gradients are also expected during periods when **[Trial mode]** is low or during periods of high wind (> 10 m/s). Temperature and humidity differences over grasslands are typically on the order of 0.5 **[Trial mode][Trial mode]** and 0.5 g **[Trial mode]**, respectively. Care must be taken in order to insure high accuracy in the measurement of the vertical differences in air temperature, humidity, and other trace species for which flux information is desirable. Bias tests (placing all sensors at the same height) are generally used to occasionally check on the accuracy of the measurements. The BREB method is also subject uncertainty from errors the measurements of **[Trial mode]** and **[Trial mode]**.

2. Modified Bowen-Ratio Method (MBR)

Similar to the BREB, the modified Bowen-ratio method (Hicks and Wesely, 1978) is not subject to all of the same constraints. With this method, a vertical flux (usually **[Trial mode]**) is determined via the eddy covariance method. Concentrations of another chemical species and air temperature are measured over two heights above the surface such that the flux of the trace species (**[Trial mode]**) is computed as **[Trial mode]** where **[Trial mode]** is the concentration difference measured over the same height interval as the air temperature. Similar to the BREB method, equality among the eddy diffusivities is assumed. Meyers et al. (1996) showed this to be the case for water vapor and carbon dioxide over a lake and forest floor of a deciduous forest. For short surfaces, in which the source/sinks are confined to a finite layer, atmospheric turbulence transports scalar entities indiscriminately and the similarity assumption is valid. This method has been successfully used to determine the fluxes of nitric acid vapor (Huebert and Robert, 1985; Meyers et al., 1989; Muller et al., 1993), ammonia (Duyzer et al., 1992), and mercury (Meyers et al., 1996; Lindberg et al., 1995). Again, the challenge is to have enough confidence in the sampling methodology to resolve the relatively small differences in concentrations between the two heights. If two different sensors are used to obtain the gradient, care must be taken to insure the bias between these two sensors is small relative to the gradient. Alternatively, a single analyzer can be used to measure the gradient by switching valves connected to intake lines for each measurement height to minimize the bias error. Although information is lost because sampling is not occurring simultaneously at each height, the error is generally small if the cycle time is less than 120 s (Meyers et al., 1996). The gradient measurement can be maximize by placing the lowest measurement level just above the surface and the upper level 2 to 5 meters above the lower level. Preferably, the eddy covariance instrumentation for **[Trial mode]** (or any other surrogate) should be placed between these to measurement levels.

3. Aerodynamic method

Only used infrequently now, the aerodynamic method assumes similarity with momentum in the flux-gradient process. Additionally, measurements at more than two heights are generally required to obtain the necessary flux-profile relationship. We will not delve into further detail of this method because of its relative infrequent use today. Instead, we refer the readers to the excellent review by Webb (1965) and Kanemasu et al. (1979) for more details and background of the aerodynamic method.

4. Integrated Horizontal Mass Flux

The integrated horizontal flux method (Beauchamp et al., 1978; Wilson et al., 1983; Wilson and Shun, 1992) is a relatively simple micrometeorological method that can be used to estimate the rate of gas transfer from the ground to the atmosphere. The method requires a small circular plot (radius **[Trial mode]** 50 m) in which the average horizontal wind speed (**[Trial mode]**) and concentration (**[Trial mode]**) profile are determined in the center of the plot in such a way that the average emission rate or upward vertical flux (**[Trial mode]**) can be determined as **[Trial mode].** Although this method cannot be used to determine fluxes *from* the atmosphere to the surface. This method has been successfully used to determine volatization rates of ammonia from grassland fertilized with urea (Wilson et al., 1982) and pesticide emission rates (Majewski et al., 1989). This method is specifically targeted for controlled experiments in which the size surface over which the fluxes are to be determined can be specified.

3. SUMMARY

Micrometeorological methods have and will continue to be an important tool for addressing many issues in agriculture that involve the soil-plant-atmosphere continuum. The challenge to researchers will be how to adapt new chemical sensors and/or sampling technologies in order to work with micrometeorological methods to obtain information on fluxes. It is envisioned that advanced sensor technologies will provide information not only on trace gas concentrations, but also on carbon and nitrogen isotopes (Harper and Sharpe, 1998) that will be useful in separating out soil and plant contributions to fluxes as well as helping to interpret bi-directional exchange.

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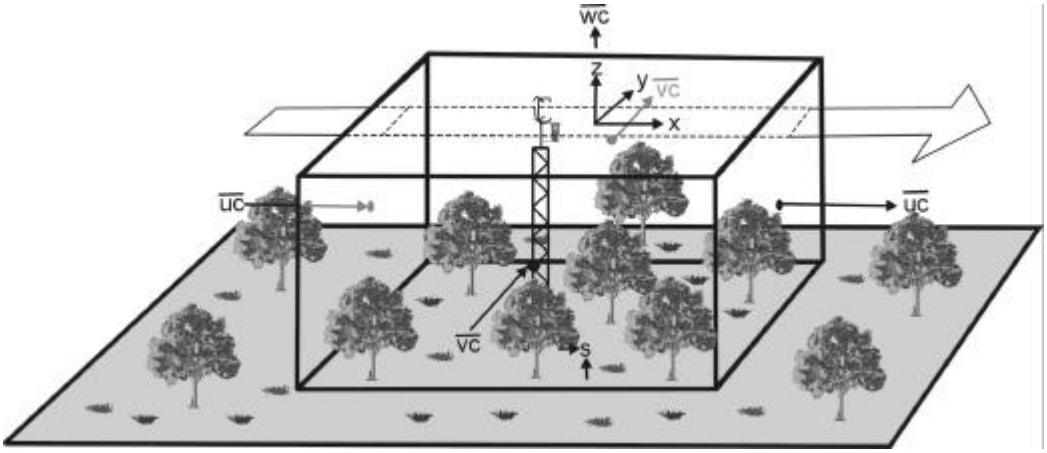
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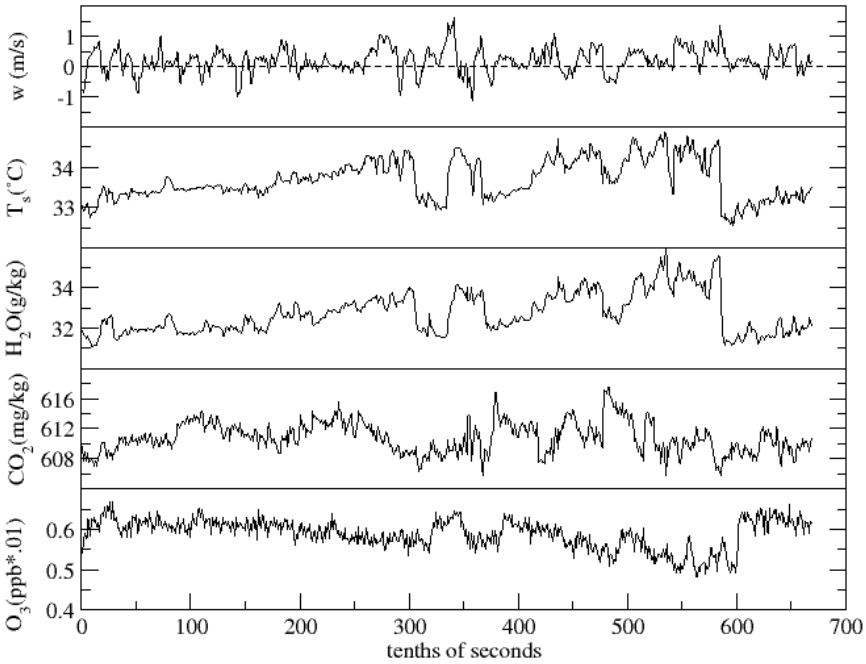
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34
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H_2O(g/kg)
34
32
CO_2(mg/kg)
616
612
608
O_3(ppb*.01)
0.6
0.5
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100
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tenths of seconds