CORRECTION OF EDDY-COVARIANCE MEASUREMENTS INCORPORATING BOTH ADVECTIVE EFFECTS AND DENSITY FLUXES

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Abstract. Equations are presented to correct eddy-covariance measurements for both fluctuations in density and non-zero mean advection, induced by convergence or divergence of flow, and spatial source/sink inhomogeneity, under steady-state and transient conditions. This correction collapses to the Webb–Pearman–Leuning expression if the mean vertical velocity is zero, and formally adds the Webb–Pearman–Leuning expression to the corrections suggested by Lee for conditions of non-zero vertical velocity and source/sink and mean scalar horizontal homogeneity. The equation requires measurement of the mean vertical gradients of the scalar concentration of interest (air temperature, humidity, CO_2) as well as an accurate estimation of the mean vertical velocity, in addition to the vertical eddy covariance of the scalar. Simple methods for the approximation of sensor tilt and complex terrain flow angle are presented, to allow estimation of non-zero mean vertical velocities. The equations are applied to data from a maize crop and a forest to give examples of when the correction is significant. In addition, a term for the thermodynamic expansion energy associated with water vapour flux is derived, which implies that the sonic temperature derived sensible heat flux will accurately include this contribution.

Keywords: Advection, Carbon dioxide, Density correction, Eddy covariance.

1. Introduction

The eddy-covariance method is a direct measure of a turbulent flux density (Swinbank, 1951, 1955) of a scalar across horizontal wind streamlines. Ecologists, agricultural, forest, micro- and bio- meteorologists and biogeochemists desire information on the exchange of material across the biosphere/atmosphere interface, rather than the flux density across streamlines. For example, the flux covariance will not equal the net ecosystem/atmosphere exchange if air is draining away or being stored under the level being measured by the eddy-covariance system. Determining how to interpret eddy-covariance measurements so that they represent the flux density of material exchanged between the atmosphere and the underlying surface is a contemporary challenge to micrometeorologists (Foken and Wichra, 1996).

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At present two theoretical lines of logic exist in the literature, for interpreting flux covariances and exchange for a scalar s. One school is derived from the classical definition of Reynolds, where the mean vertical flux is ignored and the eddy-flux is defined as:

$$F = \bar{\rho}\overline{w's'}, \tag{1}$$

where ρ is the dry air density (kg m^{-3}), s' is the scalar mixing ratio perturbation, w' is the vertical velocity perturbation, and the overbar indicates ensemble or time averaging. In this school, followed by most micrometeorologists, one rotates the coordinate system of an anemometer and forces $\bar{w}$ to zero to eliminate the mean flux term. In 1980, Webb et al. (1980) wrote a classic paper on the interpretation of eddy covariance measurements of CO_2 and other trace gas scalars with open and closed path sensors. The crux of their work was that most sensors do not measure mixing ratio (s), rather they measure the density (kg m^{-3}) of the scalar under investigation (ρ_c). In this circumstance the flux of dry air can be evaluated as a non-zero vertical velocity (Webb et al., 1980), or can be analyzed using a 'barycentric' weighting method (van Mieghem, 1949; Kramm et al., 1995). Following Webb et al. (1980), from gas laws and the simultaneous measurement of sensible and latent heat flux, they compute a mean vertical velocity and replace Equation (1) with:

$$F = \overline{w'\rho_c'} + \bar{w}\bar{\rho}_c. \tag{2}$$

Their derivation was based on absolute horizontal homogeneity, such that by continuity, if there was a vertical flux of density, a compensatory mean vertical flux of air would have to occur. Horizontal convergence was not allowed to satisfy continuity in their paradigm. Therefore, they noted, '... If the total vertical airspeed w, including $\bar{w}$ could be measured accurately, then the flux of C would not require correction, but in practice $\bar{w}$ is too small to be measured with sufficient accuracy ...' Note that in their paper C refers to a scalar. In many cases the mean vertical velocity is in fact sufficiently large to be accurately assessed, and if substituted into Equation (2), could result in unrealistically high 'corrections' much greater than the Webb et al. (1980) correction.

The other school of thought derives the eddy covariance equation from the conservation budget. In a recent paper, Lee (1998) re-visited the budget equation, see (Swinbank, 1955), for CO_2 and derived an equation that could be evaluated using conventional experimental measurement methods. He defines net ecosystem exchange of CO_2 as the sum of the eddy covariance, measured at a reference height, the storage term and an advection term. The latter term includes a mean vertical velocity ($\bar{w}$), which can be induced by local circulation or topographical drainage,

$$N_e = \overline{w'\rho_c'(h)} + \int_0^h \frac{\partial\overline{\rho_c}}{\partial t}\mathrm{d}z + \int_0^h \left[\bar{w}\frac{\partial\overline{\rho_c}}{\partial z} + \overline{\rho_c}\frac{\partial\bar{w}}{\partial z}\right]\mathrm{d}z. \tag{3}$$

Re-arrangement of the terms on the right-hand side of Equation (3) yields:

$$N_e = \overline{w'\rho_c'(h)} + \int_0^h \frac{\partial \overline{\rho_c}}{\partial t} \mathrm{d}z + \overline{w_h}(\overline{\rho_{ch}} - \langle \rho_c \rangle), \tag{4}$$

where $\langle \rho_c \rangle$ is the average scalar concentration between the ground and the canopy height h, at which there is a concentration ρ_{ch}. The term, $\bar{w}$, is defined as a mean vertical velocity measured at the canopy height. It arises from the convergence or divergence of mean horizontal wind velocity and can be non-zero. For generality, a reference height above the canopy can be substituted for the canopy height.

If for the moment, we ignore the storage of the scalar, we arrive at the following expression:

$$N_e = \overline{w'\rho_c'(h)} + \overline{w_h}(\overline{\rho_{ch}} - \langle \rho_c \rangle). \tag{5}$$

The problem we face here, by contrasting Equations (2) and (5), is two potentially competing versions of a vertical velocity correction to assess net ecosystem exchange of a scalar. Is there an inconsistency between the mean vertical velocity applied by Webb et al. (1980) theory versus that applied by Lee (1998)? For some, there is still the question as to whether one uses the Webb et al. (1980) correction and vertical advection terms together, or if one uses only the larger of the two (Laubach and Teichmann, 1999).

In this paper, we re-derive the budget equations for scalars and attempt to reconcile perceived differences between the two schools of thought for evaluating turbulent fluxes. We derive a new thermodynamic energy budget correction term associated with the latent energy flux. We also discuss practical methods for assisting in assessing non-zero, mean vertical velocities, and discuss equations for transient conditions when moving averages may be indicated for certain terms. Several examples of data analysis are given.

2. Conservation of Mass

The conservation of mass of species c is

$$\frac{\partial \rho_c}{\partial t} + \frac{\partial (u_i \rho_c)}{\partial x_i} = S_c, \tag{6}$$

where u_i is the velocity component in direction i, and S_c is the Eulerian source term for the scalar, in units of kg m^{-3} s^{-1}. After Reynolds decomposition and averaging, this equation becomes,

$$\frac{\partial \bar{\rho}_c}{\partial t} + \frac{\partial (\bar{u}_i \bar{\rho}_c)}{\partial x_i} + \frac{\partial (\overline{u_i' \rho_c'})}{\partial x_i} = \bar{S}_c \tag{7}$$

using ‘Einstein’ notation where the repeated indices i indicate summation over the values i = 1, 2 and 3.

If one switches to meteorological notation where $u = u_1$, $v = u_2$, and $w = u_3$, and carries out scale analysis,

$$\frac{\partial(\bar{u}\bar{\rho}_c)}{\partial x} \gg \frac{\partial(\overline{u'\rho_c'})}{\partial x} \tag{8}$$

and if one also assumes that the mean cross-wind velocity is zero (i.e., aligning the coordinate system along the mean wind), and that the cross-wind gradients of mean quantities are zero, the following equation results:

$$\frac{\partial \bar{\rho}_c}{\partial t} + \bar{\rho}_c \frac{\partial \bar{u}}{\partial x} + \bar{u}\frac{\partial(\bar{\rho}_c)}{\partial x} + \bar{\rho}_c \frac{\partial \bar{w}}{\partial z} + \bar{w}\frac{\partial(\bar{\rho}_c)}{\partial z} + \frac{\partial(\overline{w'\rho_c'})}{\partial z} = \bar{S}_c, \tag{9}$$

which implies that to estimate the source or sink of a scalar entity, horizontal as well as vertical gradients of mean quantities must be measured, in addition to eddy-covariance in the vertical direction. These requirements can be reduced by invoking the equation of continuity for dry air, written to include the possibility of spatial and temporal changes in the mass density, with similar assumptions to those above, i.e., the gradient of the longitudinal eddy-covariance is much smaller than the gradient of the mean advective terms, and the cross-wind mean velocity is zero, as is the gradient of the covariances.

It is instructive to note at this point that Webb et al. (1980) ignored the horizontal advection terms of Equation (9), with the stated assumption of complete horizontal homogeneity, but accounted carefully for the air density flux term in the vertical. In contrast to Webb et al. (1980), Lee (1998) assumed that incompressible flow existed such that the air density could be assumed as a constant, but allowed a form of vertical advection, arising from a mean vertical velocity, which near a surface could only occur with a horizontal gradient in the mean horizontal velocity field. We contend that in many field conditions, most or all of the terms in Equation (9) must be considered for completeness.

We introduce here the notion that two major forms of horizontal homogeneity can be identified. The first is horizontal scalar homogeneity, where the mean horizontal gradients of scalars and horizontal covariances are negligible. This implies source element horizontal homogeneity, such as a homogeneous plant canopy, when averaged over the ‘footprint’ of the sensors. Under these conditions, mean horizontal gradients of the longitudinal velocity are still possible, as one might expect under typical mesoscale, synoptic scale, or other circulations, so that the mean velocity gradient times the mean scalar terms of Equation (9) would be important but the horizontal mean scalar term could be ignored.

One might consider that even under source homogeneity, that there would be some non-zero horizontal gradients of the scalars (Finnigan, 1999). While this is likely, these gradients are generally observed to be much smaller than vertical gradients (which are closely linked to the vertical inhomogeneity of source strength). For example, it is common for vertical air temperature gradients to be on the order of several K per 10 m interval within tall plant canopies and at least 10 times that gradient for short crop canopies. The equivalent horizontal gradients, for homogeneous source-plant canopies, are usually less than several K over km distances (Lee, 1998, 1999). Even with the mean horizontal velocity typically being an order of magnitude or two greater than the mean vertical velocity, this would still imply that frequently the horizontal mean advection term would be negligible or at least smaller than the vertical term.

Discussions by Lee (1998, 1999) and Finnigan (1999) consider these issues. It is certainly possible that in certain cases, even with horizontal source homogeneity, significant horizontal scalar gradients could occur that would necessitate retaining the associated terms, and imply that the horizontal gradients would have to be estimated by measurement or reliable modelling. During nocturnal drainage, under very stable conditions and low levels of turbulence, horizontal scalar gradients could yield dominant advection terms. Therefore, under such conditions, horizontal advection should be directly measured, even if difficult to accomplish.

The second form, complete horizontal homogeneity, trivially assumes that all horizontal gradients of the means and covariances of the velocity field, are zero. This is typically the assumption made when eddy covariance is used to estimate ecosystem exchange, such as the assumptions used in Webb et al. (1980). We will discuss in detail only the former, i.e., that the horizontal scalar fields are horizontally homogeneous, but that some horizontal gradients in the velocity field can occur.

Following Webb et al. (1980), we substitute the product of the mixing ratio of species $c(s)$ and the dry air density for the species mass density, ρ_c, such that the effects of dry air density changes caused by perturbations in temperature and water vapor (for moist air) can be separated from the effects of mixing ratio perturbations. We then derive Reynolds averaged approximations for use in the expressions:

$$\rho_c = s\rho, \tag{10}$$

$$\overline{\rho_c} = \bar{s}\bar{\rho} + \overline{s'\rho'} \approx \bar{s}\bar{\rho}. \tag{11}$$

The time derivative of Equation (11) is,

$$\frac{\partial \overline{\rho_c}}{\partial t} = \frac{\partial \bar{s}}{\partial t}\bar{\rho} + \frac{\partial \bar{\rho}}{\partial t}\bar{s} + \frac{\partial \overline{s'\rho'}}{\partial t} \approx \frac{\partial \bar{s}}{\partial t}\bar{\rho} + \frac{\partial \bar{\rho}}{\partial t}\bar{s}, \tag{12}$$

The substitutions are then placed into Equation (9), resulting in:

$$\bar{\rho}\frac{\partial \bar{s}}{\partial t} + \bar{s}\frac{\partial \bar{\rho}}{\partial t} + \bar{s}\bar{\rho}\frac{\partial \bar{u}}{\partial x} + \bar{s}\bar{u}\frac{\partial \bar{\rho}}{\partial x} + \bar{u}\bar{\rho}\frac{\partial \bar{s}}{\partial x} + \bar{s}\bar{\rho}\frac{\partial \bar{w}}{\partial z} + \bar{w}\bar{\rho}\frac{\partial \bar{s}}{\partial z}$$
$$+\bar{w}\bar{s}\frac{\partial \bar{\rho}}{\partial z} + \frac{\partial \overline{w'\rho_c'}}{\partial z} = S_c. \tag{13}$$

Next, continuity is used:

$$\frac{\partial \rho}{\partial t} + \frac{\partial (u_i \rho)}{\partial x_i} = 0, \tag{14}$$

which, with Reynolds decomposition, averaging and some assumptions, leads to

$$\bar{\rho}\frac{\partial \bar{u}}{\partial x} = -\bar{w}\frac{\partial \bar{\rho}}{\partial z} - \bar{\rho}\frac{\partial \bar{w}}{\partial z} - \frac{\partial (\overline{w'\rho'})}{\partial z} - \bar{u}\frac{\partial \bar{\rho}}{\partial x} - \frac{\partial \bar{\rho}}{\partial t}. \tag{15}$$

The last form of continuity is substituted into Equation (13), and under conditions when the horizontal gradient of the scalar can be ignored, the equation becomes:

$$\bar{\rho}\frac{\partial \bar{s}}{\partial t} + \bar{s}\frac{\partial \bar{\rho}}{\partial t} + \bar{s}\left(-\bar{w}\frac{\partial \bar{\rho}}{\partial z} - \bar{\rho}\frac{\partial \bar{w}}{\partial z} - \frac{\partial (\overline{w'\rho'})}{\partial z} - \bar{u}\frac{\partial \bar{\rho}}{\partial x} - \frac{\partial \bar{\rho}}{\partial t}\right) + \bar{s}\bar{u}\frac{\partial \bar{\rho}}{\partial x}$$

$$+\bar{u}\bar{\rho}\frac{\partial \bar{s}}{\partial x} + \bar{s}\bar{\rho}\frac{\partial \bar{w}}{\partial z} + \bar{w}\bar{\rho}\frac{\partial \bar{s}}{\partial z} + \bar{w}\bar{s}\frac{\partial \bar{\rho}}{\partial z} + \frac{\partial (\overline{w'\rho_c'})}{\partial z} = \bar{S}_c \tag{16}$$

with several terms canceling each other out, and assuming horizontal homogeneity of the scalar s,

$$\bar{\rho}\frac{\partial \bar{s}}{\partial t} - \bar{s}\frac{\partial (\overline{w'\rho'})}{\partial z} + \bar{w}\bar{\rho}\frac{\partial \bar{s}}{\partial z} + \frac{\partial (\overline{w'\rho_c'})}{\partial z} = \bar{S}_c. \tag{17}$$

With the substitution for ρ' for temperature and moisture effects under assumed constant pressure, from Webb et al. (1980), the equation becomes,

$$\bar{\rho}\frac{\partial \bar{s}}{\partial t} + \bar{s}\frac{\partial}{\partial z}\left[\frac{\bar{\rho}}{\bar{T}}(1+\mu\sigma)(\overline{w'T'}) + \mu(\overline{w'\rho_v'})\right] + \bar{w}\bar{\rho}\frac{\partial \bar{s}}{\partial z} + \frac{\partial (\overline{w'\rho_c'})}{\partial z} = \bar{S}_c, \tag{18}$$

where ρ_v is the absolute humidity,

$$\mu \equiv \frac{m_a}{m_v} \quad \text{and} \quad \sigma \equiv \frac{\bar{\rho}_v}{\bar{\rho}},$$

where m_a is the molecular mass of dry air, and m_v is the molecular mass of water. When horizontal heterogeneity is apparent, the horizontal gradients of the scalar would also have to be included in Equation (18). The issues of integrating this equation are discussed later in this paper and in Appendix A.

This expression is very similar to that of Webb et al. (1980), except that now a mean advection term has appeared (Lee, 1998), in the form of the measured mean vertical velocity times some mean variables (term 3 on the left hand side), and the equation is in differential form, stemming from its derivation from the conservation of mass. In contrast to the interpretation of Webb et al. (1980), that the vertical air density flux term (term 2 on the left hand side), represents a vertical velocity, in our interpretation the density flux emerges from continuity as a component of both horizontal and vertical advection of dry air. Webb et al. (1980) advised that if accurate measurement of the mean vertical wind speed were available, that the correction term (term 2) would not be necessary. This was entirely because of their assumption that there was no horizontal advection possible; in their paradigm, the *only* mean vertical velocity possible would be that exactly balancing the eddy-covariance of dry air density.

Our derivation shows that when mean vertical velocities are measured, usually at magnitudes somewhat greater (by an order of magnitude or more for forests) than that implied by Webb et al. (1980) correction, one should *add* the mean advective contribution to the Webb et al. (1980) correction arising from the mean air density eddy-covariance. If the mean advection term is the opposite sign of the Webb et al. (1980) correction, the total correction could be *less* than the Webb et al. (1980) correction alone. As discussed by Webb et al. (1980) if the eddy covariance is performed on a scalar concentration expressed as a specific mass ratio, such as from a 'closed-path' infrared gas analyzer (IRGA), then the mean air density eddy-covariance is canceled by the closed-path IRGA conversion of mass concentration s to the mass per unit volume ρ_c. The mean vertical velocity term would remain. If the measured mean vertical velocity is zero, with horizontal homogeneity of the mean scalars, our expression collapses to that of Webb et al. (1980) after integration.

3. Importance of the Vertical Velocity Compared to the Density Correction

When the mean vertical velocity is non-zero, the effects of mean advection can be significant (Lee, 1998). Hypothetical calculations, plotted in Figure 1, show that the advection correction can be as significant as the eddy covariance of density correction at mean vertical velocities of less than 0.05 m s^{-1}, which are not rare at typical measurement heights above plant canopies, especially forests and even shorter crop canopies. Because the correction factor depends linearly on the mean scalar concentration gradient and the mean vertical velocity, individual sites, days, and times can have very large advective corrections, or negligible corrections.

Our derivation up to now has been for a differential equation at a point in space, with no physical volume. In practice, for the source/sink of non-zero size, we must formally integrate Equation (18) over a volume in the x, y, and z direction, and divide by the horizontal volume dimensions, to yield the exchange expressed per horizontal surface area of the volume (flux density). Under horizontal homogeneity of the source/sink, the integration and division by the horizontal dimensions cancel, leaving only the vertical integration to be performed. For the vertical integration evaluated at some height z_1, Equation (18) may be integrated with limits of 0 to z_1, similar to the method used by Lee (1998). All of the terms will generally be height

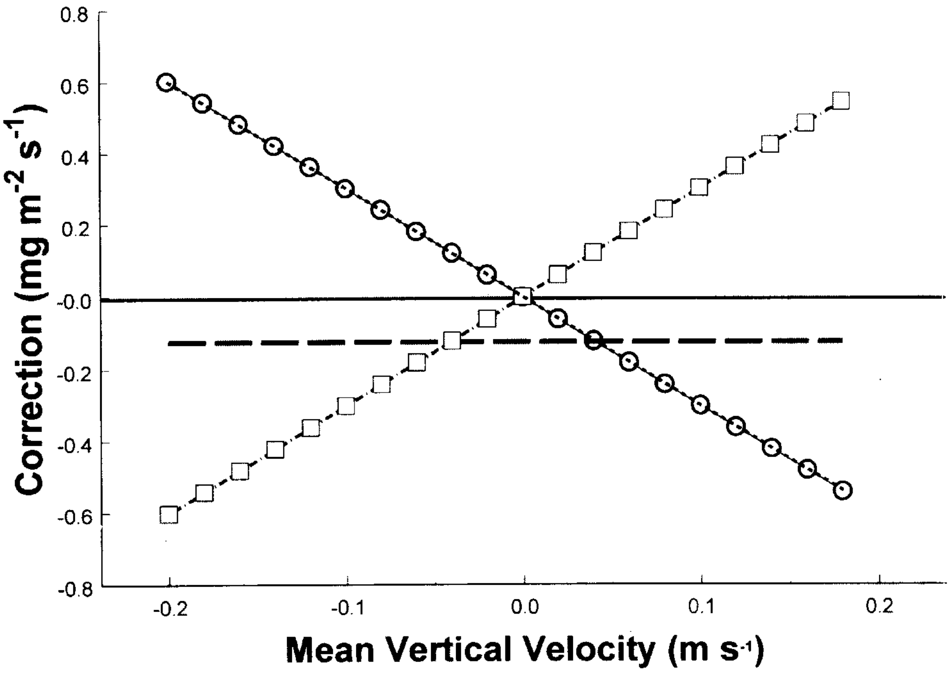

Figure 1. Correction of scalar flux density (CO_2) as a function of mean vertical velocity. Squares show the correction for soil concentration 5 ppm less than that at the measurement height; circles a soil concentration 5 ppm greater than that at the measurement height, and the dotted line indicates the dry Webb et al. (1980) correction at -100 W m^{-2}.

dependent, and without a simple analytical form within a plant canopy air space, so it is easiest use a numerical integration method (see Appendix A). Under horizontal heterogeneity of the scalar, the appropriate terms in Equation (13) would have to be included in the derivation (implying measurements or estimates of the mean scalar horizontal gradients would have to be made), and then would be subject to numerical vertical integration.

4. Practical Methods for Terrain-Following Flow and Sensor Tilt

Traditionally, many researchers have assumed the mean vertical velocity must be zero for each data period (usually 20 to 30 min), and developed equations to trigonometrically rotate axes. This stemmed from typically short-term experiments, when extensive data from different wind directions did not exist. Generally, in such experiments, wind from only a narrow range of directions would be accepted to assure reasonable fetch and minimize probe mounting and tower/mast interference. Therefore, each data period was assumed to yield flows which followed terrain exactly, so in terrain-following coordinates no vertical wind speed was possible. Under such an assumption, flux correction for any real non-zero w is impossible.

The development of longer-term sonic anemometer measurements allows usage of a practical method to estimate the tilt and terrain following flow (this method, as with traditional axis rotation, cannot distinguish between terrain-following flow and sensor tilt). Lee (1998) suggests that the mean vertical velocity is 'random', which we interpret to mean that on the average, it is zero over a 'long' period of time. He suggests the following equation type,

$$\bar{w}_e = \bar{w} - a(\eta) - b(\eta)\bar{u}, \tag{19}$$

where w_e is estimated and the coefficients a and b are functions of the azimuthal angle η.

Here, we identify three possible general methods, one of which includes Lee's (1998) specific method, and note that there have been several discussions, such as by Bernstein (1966) and Smith et al. (1985). In method (1), one arrives at some relationship between the vertical transform angle θ and the azimuthal transform angle η (see Figures 2–4). This could include something similar to Lee's method, because the vertical transform angle is calculated as the arctangent of the mean vertical velocity divided by the mean horizontal velocity. Or, a polynomial could be fit to the relationship, or a sinusoidal function. If the fit is close to a sinusoidal function (see Figures 2 and 3) this method is identical conceptually to general method (2), which assumes that the terrain has a uniform tilt, so that the combination of instrument tilt and the terrain produce the equivalent of the instrument over a tilted plane. If the fit is not sinusoidal, the terrain (more rigorously, the mean three-dimensional flow patterns) does not exhibit a uniform tilt, for example, an area with several small hills and valleys.

In one form of method (2), used by Kondo and Sato (1982) and Mahrt et al. (1996), with the assumption that the surface is uniformly tilted in respect to the sensor (equivalent to the sensor being uniformly tilted with respect to the surface), a two-dimensional linear regression of mean vertical velocity is taken on the mean horizontal velocity components in the x and y directions, before any angle transformation, for data gathered over a period ranging from days to months. A tilted plane results from the regression (see Figure 5). This plane can easily be visualized as the tilted surface that the sensor 'sees'. The regression equations yield a quick way to approximate the portion of the measured mean vertical velocity which results from the combined effects of terrain-following flow and instrument tilt. For more complex terrain, or to account for sensor probe and mast/tower influences, a non-linear surface regression can be used.

In method (3), the terrain surrounding an instrument is carefully surveyed, and the instrument is carefully leveled. Computations can be made to trigonometrically estimate correction angles for any horizontal wind direction. This method will not be tested in this manuscript, because it is trivial conceptually, but is difficult to carry out in practice when frequently the instrument tilt itself can be a degree or more, even after careful leveling, and this may vary with periodic maintenance to nearby equipment on a tower or mast.

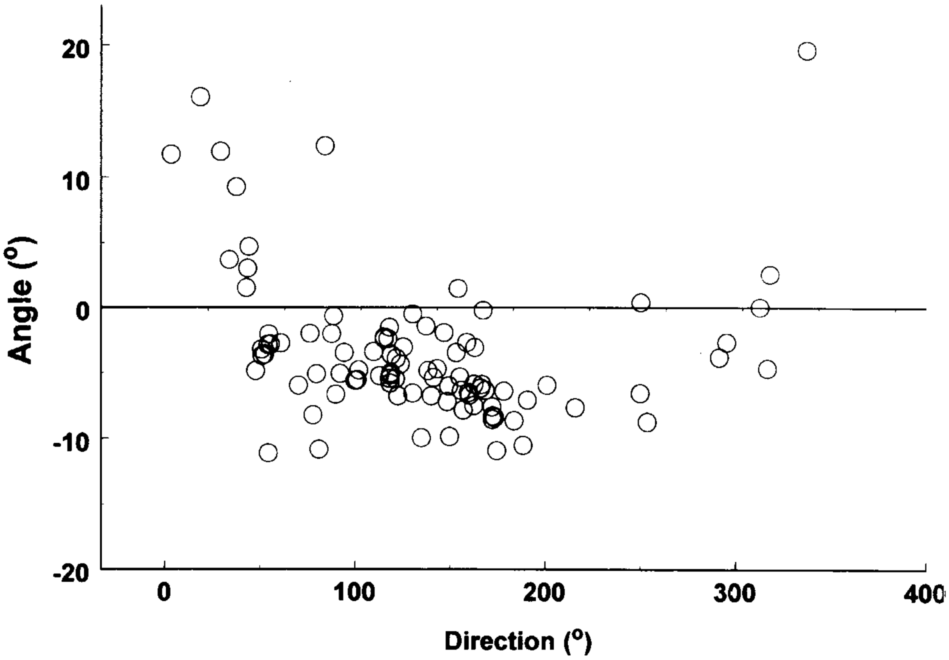

Figure 2. Vertical transform angle as a function of horizontal wind direction of the Wind River Canopy Crane Facility AmeriFlux site.

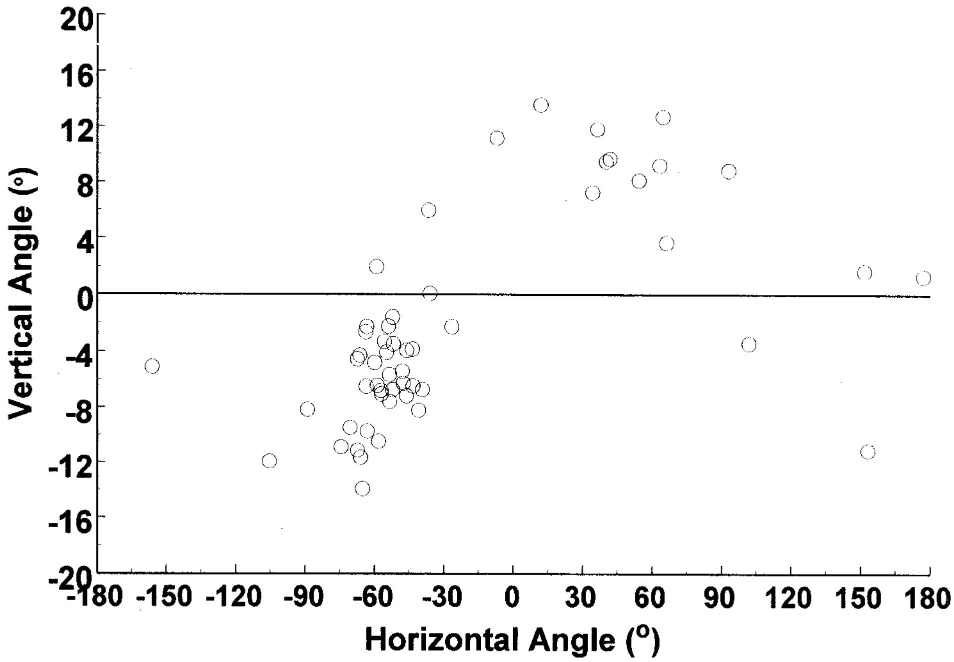

Figure 3. Vertical transform angle as a function of horizontal wind direction for the measurements taken at 3 m under a Walnut canopy in Winters, California, 1989.

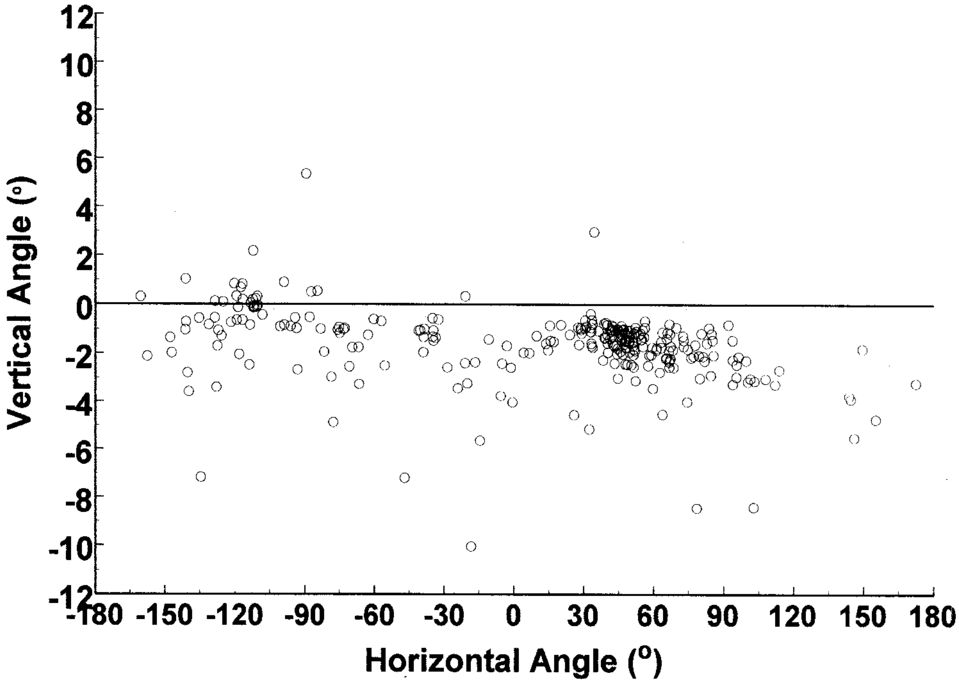

Figure 4. Vertical transform angle as a function of horizontal wind direction for the measurements taken at 2.6 m over a maize canopy in Davis, California, 1988.

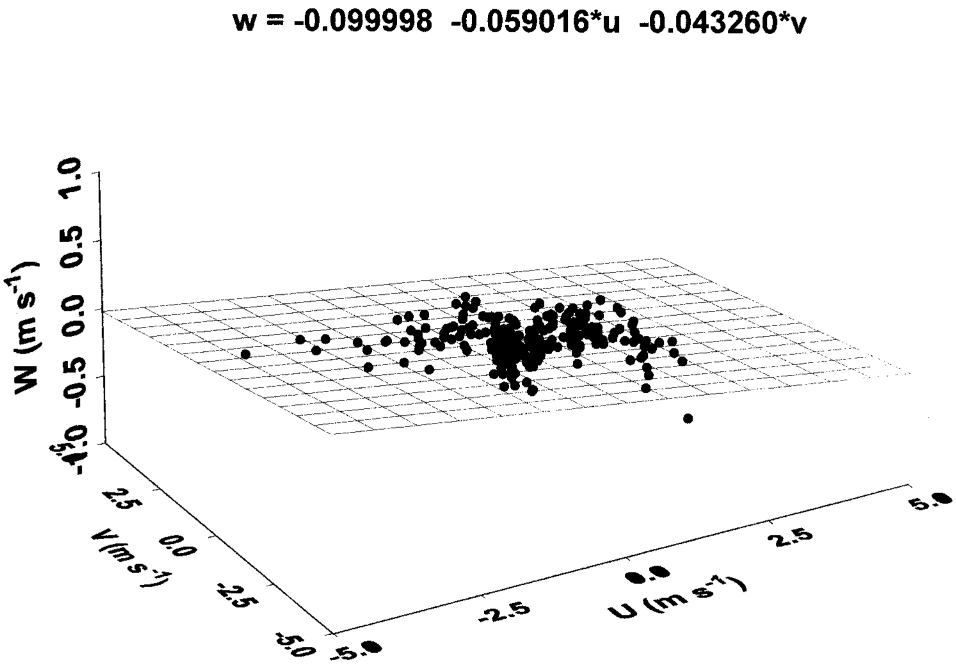

Figure 5. Two-dimensional plane regression for the same data displayed as angles in Figure 2.

Method (2) is a quick way to apply corrections to vertical velocity data, and in addition, tends to emphasize the stronger winds that frequently will contribute more to exchange estimates than lower winds. At lower wind speeds, the vertical transform angles may vary greatly, influencing a methodology such as covered in method (1), where angles are used to form relationships instead of velocity data. Method (1) does have the ability to correct for asymmetric sensor probe and mast/tower influences, whereas the linear regression form of method (2) does not. We should note that method (2) is based on the assumptions that the mean vertical velocity does have a consistent planar relationship to the mean horizontal winds, as would happen for an ideal sensor tilted in respect to an underlying surface. Synoptic and mesoscale flows in certain regions and terrain may violate this assumption, with mean convergent or divergent flows creating other relationships between vertical and horizontal velocities.

Finally, because orthogonal axes rotations have no effect on the absolute velocity vectors or the scalars, the exchange equations presented in previous sections could actually involve *any* arbitrary rotation, so long as all mean and covariant terms are included, with special attention paid to the mean horizontal and vertical advective terms which must be estimated based on the gradients and velocity components in the transformed axes frame. For modelling and analysis purposes, however, many researchers would choose a pseudo-streamline or terrain following coordinate system, which should match the tilt estimation methods outlined above. The assumptions concerning homogeneity have to be carefully formulated in regards to axes rotation. For example, from the assumption that scalar gradients are mainly created by gradients in source/sink strength, one would conclude that a predominantly ‘vertical’ gradient could appear as a ‘horizontal’ gradient in a radically transformed axis system. ‘Horizontal’ homogeneity (in the new, transformed axes) could no longer be assumed, but ‘vertical’ homogeneity could be assumed.

5. Applications with Experimental Data

Two examples, each for a sample day, are given in this paper, from temperature data taken above a 2.6-m high maize crop and CO_2 data taken above a 24-m high deciduous forest, for evaluation of the importance of the vertical advection term. The experimental set up from the maize crop was described in greater detail in Paw U et al. (1992). Details concerning the Walker Branch forest site in Oak Ridge, TN, and experimental set up, can be found in Meyers and Baldocchi (1993) and Baldocchi and Vogel (1996).

During the day, over a maize crop, the eddy-covariance term accounts for most of the sensible heat exchange, although the advection can be substantial although still smaller than the eddy covariance term (Figure 6). At night, the advection term is a significant contributor to the sensible heat exchange. The transient storage term is negligible throughout the 24-hour day. The eddy covariance, traditionally the

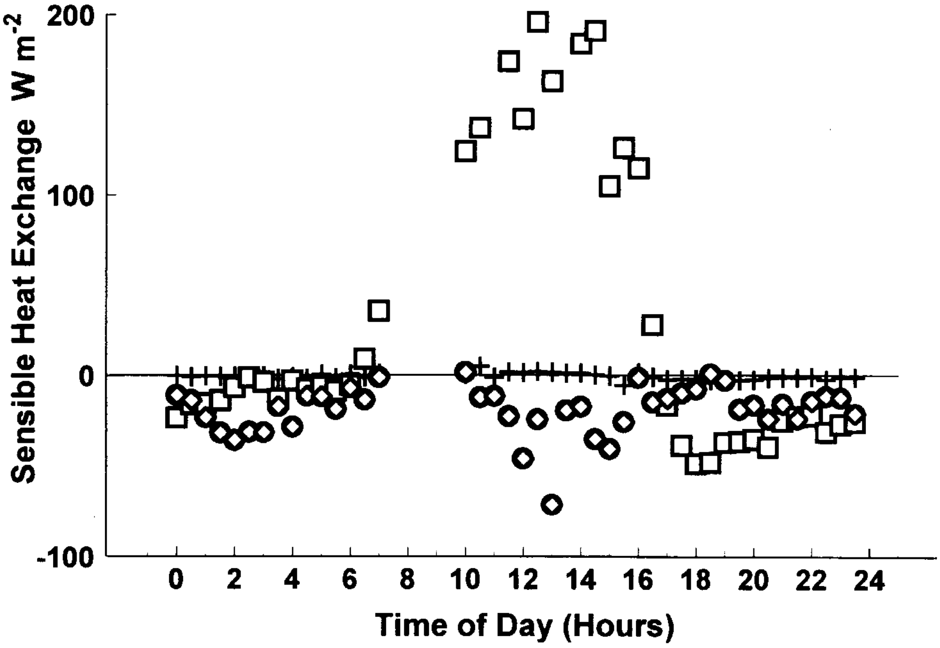

Figure 6. Contributions of terms in the sensible heat exchange equation, for a maize canopy in Davis, California. The pluses represent the transient storage term, the circles advection, and the squares the vertical eddy-covariance term.

only term used for exchange calculations, is not symmetric (about noon) at night and early morning, in other words the negative sensible heat is of a lower magnitude in the early morning than it is in the late evening (Figure 7). The expectation is that the available energy ($Rn - G$) was approximately equal during both periods, leading to a more symmetrical time course for the sensible heat flux (because the latent energy term was negligible during these times). The full exchange equation yields a more reasonable time course, with the early morning and night values approximately equal. During the day, both exchange methods show more scatter than during the night.

At the Walker Branch deciduous forest, the time course of the carbon dioxide flux components on day 293 (1997) also shows that the vertical advection term can be an important term at night, when it is positive, while the transient term is generally smaller at night but can be a significant term in the morning (Figure 8). Eddy covariance is the dominant term during the day, under photosynthetic conditions, but is not as important at night, when it can even be of the wrong sign (a biometeorologically unrealistic sink of carbon with no light). The exact cause of the unrealistic eddy-covariance values such errors could stem from several processes, but we did not determine which ones. The Webb et al. (1980) term is significant during the day, but insignificant at night, because the main contributors to it, sensible heat and latent energy, are low in magnitude.

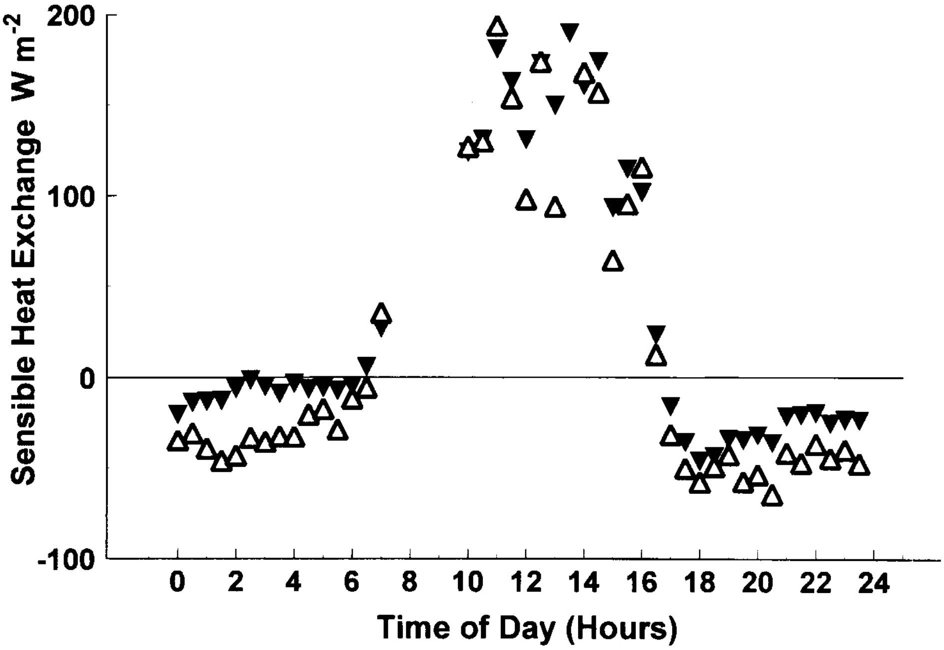

Figure 7. Sensible heat exchange, for a maize canopy in Davis, California. Open triangles represent entire corrected exchange equation. Inverted solid triangles represent eddy covariance.

A similar pattern for the components occurs on a different day in 1997, DOY 273 (Figure 9). The vertical advection term is important at night again, and the storage term important at various times. The Webb et al. (1980) correction is significant during the day. Eddy covariance is the major component in the day, but is not as dominating at night.

The relative contributions of each component, normalized to the net ecosystem exchange (NEE), are shown in Figures 10 and 11. A value of one means that the component represents all of the NEE. Advection is clearly important at night, and eddy-covariance overestimates the carbon sink overestimates during the day. The Webb et al. (1980) expression corrects this overestimate during the day, but is negligible during the night.

It is important to notice the difference in the importance of components, between sensible heat and CO_2 exchange. In CO_2 , vertical advection is important at night, with eddy covariance and the Webb et al. (1980) correction important during the day. Advection is important during both the day and night for sensible heat. Storage is important for carbon, in a forest, but for sensible heat in a crop, it is unimportant. One can speculate that nocturnal latent energy flux density could be less sensitive to advective errors, because it is frequently small, leading to the possibility that the energy budget could appear to approximately close without

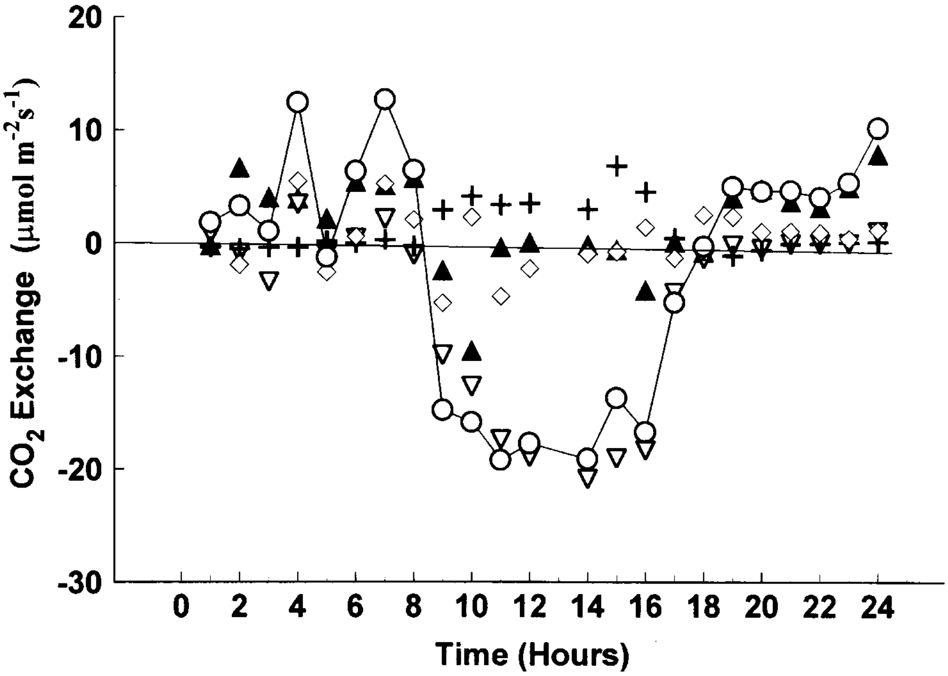

Figure 8. Components of the net ecosystem exchange (NEE) of CO_2 over a deciduous forest, day 293. The circles represent the NEE, the inverted open triangles, the eddy covariance, the solid triangles, the vertical advection, the diamonds, the transient storage term, and the pluses, the Webb et al. (1980) correction.

including the advective components for a particular experiment, while the CO_2 estimates could still contain large errors.

6. Transient Conditions

Many times in the day are characterized by transient conditions. Under such transiency, the formal Reynolds analysis appears not to be strictly valid for conservation equations, mainly because of the appearance of an block averaged term within a time derivative, such that the time derivative of the block average cannot be meaningfully performed (see Equation (18), first term). If a moving average is used in its place (Bernstein, 1966, 1970), then the conservation equations and Reynolds decomposition can be modified, although sometimes terms may not drop out. It should be noted that the time derivative of the block averaged term is formally valid (it will be zero), and the transient term becomes only the block mean of the derivative of the perturbation term. However, in many cases the mean of a time derivative is more accurately measured by slow response sensors, which exhibit

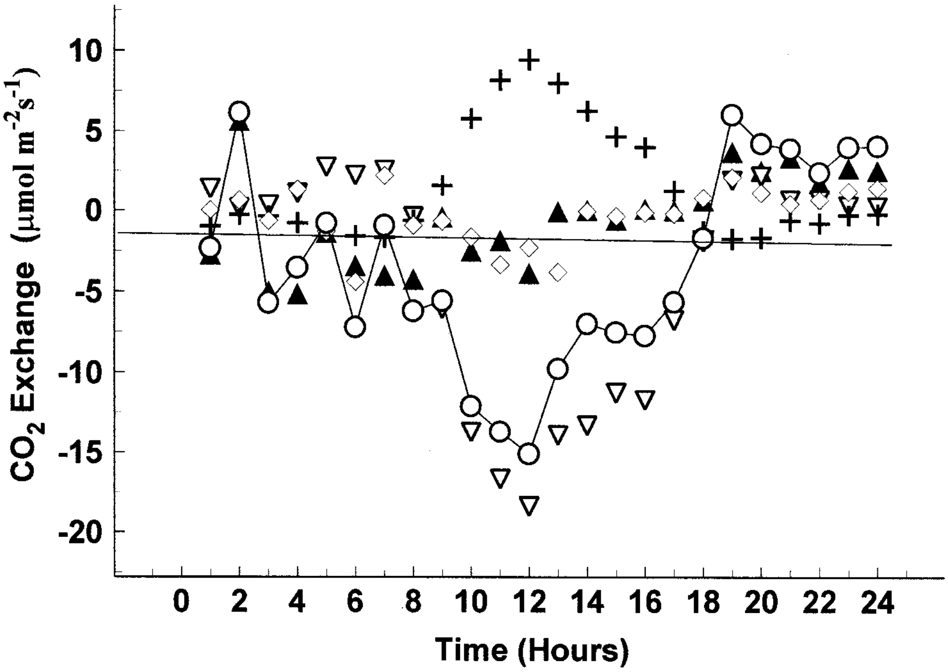

Figure 9. Components of the net ecosystem exchange (NEE) of CO_2 over a deciduous forest, day of year 273. The circles represent the NEE, the inverted open triangles, the eddy covariance, the solid triangles, the vertical advection, the diamonds, the transient storage term, and the pluses, the Webb et al. (1980) correction.

less drift than fast-response sensors. Because slow-response sensors represent a form of moving average filtering, the technique will be briefly discussed below.

The moving average of variable $s(t)$ is formally described as,

$$[s] \equiv \frac{1}{\tau}\int_{t-\tau/2}^{t+\tau/2} s(\xi)\,\mathrm{d}\xi, \tag{20}$$

where τ is the averaging time, the brackets indicate a moving average, and the result is a function of the midpoint time t. For the time derivative of s, for example,

$$\frac{\partial s(t)}{\partial t} = \frac{\partial [s(t)]}{\partial t} + \frac{\partial s(t)''}{\partial t} \tag{21}$$

can be used, because $[s(t)]$ is a function of time; note that the double primes indicate a perturbation from the moving average. However, if the moving average is used in Reynolds decomposition,

$$w = [w] + w'', \tag{22a}$$

$$\rho = [\rho] + \rho'', \tag{22b}$$

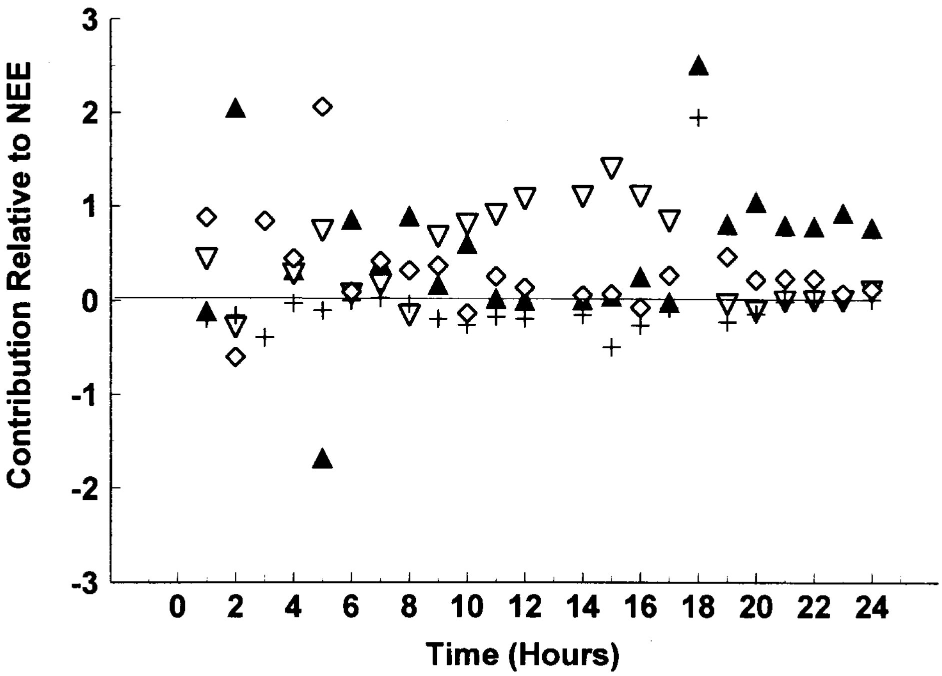

Figure 10. Relative contributions of the components of NEE for day of year 293. The symbols are the same as used in Figures 8 and 9.

$$w\rho = [w][\rho] + w''[\rho] + [w]\rho'' + w''\rho'' \tag{22c}$$

and then the moving average applied to Equation (22c), in a similar manner to what is done in conventional decomposition with block averaging,

$$[w\rho] = [[w][\rho]] + [w''[\rho]] + [[w]\rho''] + [w''\rho''], \tag{22d}$$

then the average of the products of the mean and perturbation terms are not necessarily zero. For example, following Bernstein's (1966) experimental calculations or using order-of-magnitude analysis, one can deduce that the product $[w''[\rho]]$ could be equal or greater in magnitude to the product of the means, $[[w][\rho]]$. We also confirmed this deduction with some calculations using the maize experiment data. This would add extra terms to the conservation equations and make them cumbersome to analyze.

However, there is no reason that a block average of the equations cannot be performed after an initial Reynolds decomposition using moving averages. This will retain the ability to take a time derivative of the slow response based scalar concentration, but as shown below, simplifies the covariance calculations. The

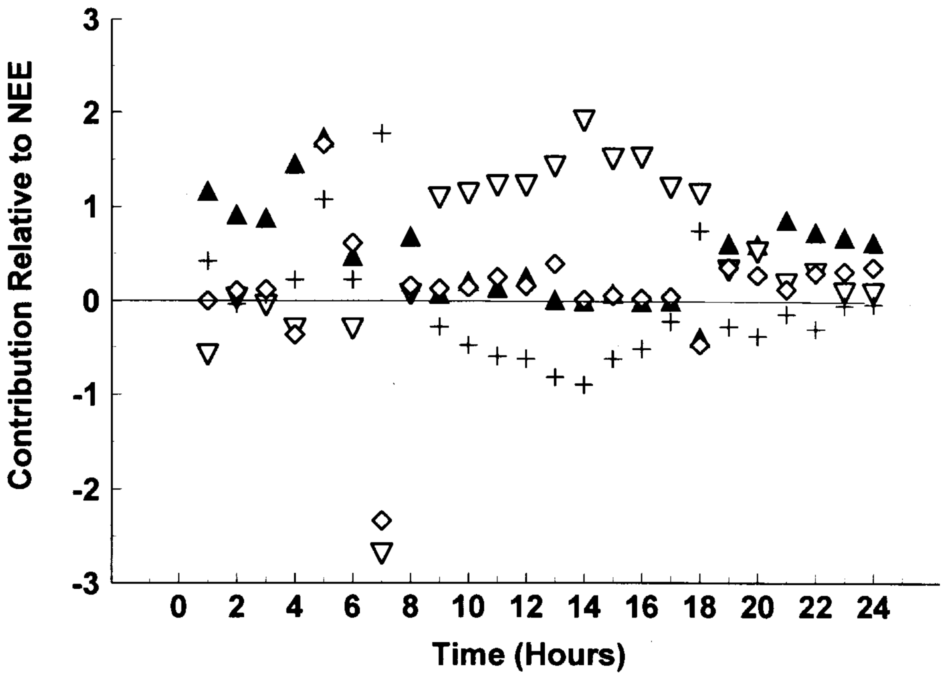

Figure 11. Relative contributions of the components of NEE for day of year 273. The symbols are the same as used in Figures 8 and 9.

block-averaged equation for the transient term (using moving-average Reynolds decomposition) is,

$$\overline{\frac{\partial s(t)}{\partial t}} = \overline{\frac{\partial [s(t)]}{\partial t}} + \overline{\frac{\partial s(t)''}{\partial t}}, \tag{23}$$

where perhaps if the perturbations of the scalar concentrations are small compared to the mean, which is frequently true, the second term on the right hand side may be ignored (see Appendix B). It should be noted here that if a forward-difference operator is used for the left hand side of Equation (23), then it can be shown that the time derivative of $s(t)$ is also equal to the starting value for s and the ending value for s divided by the block averaging interval (see Appendix B). Random fluctuations in s may insert error using this method.

For the covariances, it appears initially that all four terms must be retained:

$$\overline{w\rho} = \overline{[w][\rho]} + \overline{w''[\rho]} + \overline{[w]\rho''} + \overline{w''\rho''}. \tag{24}$$

But it should be recalled that Equation (24) must be equal to $\overline{w\rho}$ evaluated with conventional Reynolds decomposition, such that

$$\overline{w\rho} = \overline{[w][\rho]} + \overline{w''[\rho]} + \overline{[w]\rho''} + \overline{w''\rho''} = \bar{w}\bar{\rho} + \overline{w'\rho'}. \tag{25}$$

Therefore, so long as both the mean and the product of perturbation terms are retained in block-averaged conservation equations, they may be used instead of moving averaged terms. Although one may formally calculate the moving average terms as above, the sum of all four moving average terms must always be equal to the sum of the two retained block averaged terms. The only place one may need to retain the moving average operation is within the time derivative. The conservation equation then becomes,

$$\overline{[\rho]\frac{\partial [s]}{\partial t}} + \bar{s}\left[\frac{\bar{\rho}}{\bar{T}}(1+\mu\sigma)\frac{\partial(\overline{w'T'})}{\partial z} + \mu\frac{\partial(\overline{w'\rho_v'})}{\partial z}\right] + \bar{w}\bar{\rho}\frac{\partial \bar{s}}{\partial z} + \frac{\partial(\overline{w'\rho_c'})}{\partial z} = \bar{S}_c. \tag{26}$$

7. Thermodynamic Considerations

When the Webb et al. (1980) expression is used, one must assume that pressure is constant. The fact that the total static pressure perturbations in the atmosphere are usually much smaller than the vapour pressure perturbations supports the hypothesis that the total pressure at some distance from the evaporating surfaces is constant with respect to vapour pressure fluctuations. Therefore, it is then implied that the moistened air had to perform work to achieve constant pressure, i.e., the parcels had to expand to decrease their density, against an approximately constant pressure. It follows that energy exchange under non-zero latent energy conditions should include not only the sensible heat, but the equivalent amount of energy derived from the surface and air above the surface, to expand the air when moisture is added (or conversely, contract the air when atmospheric moisture is lost to surfaces). This amount of energy (work of expansion) is approximately 7.6% of the latent energy term (Appendix C).

The eddy covariance of sonic temperature, which is approximately that of virtual temperature, approximates this extra thermodynamic energy. The virtual temperature covariance is given by

$$\overline{w'T_v'} \approx \overline{w'T'}(1+0.61\bar{r}) + 0.61\bar{T}\,\overline{w'r'} \approx \overline{w'T'} + 0.61\bar{T}\,\overline{w'r'}, \tag{27}$$

where r is the mixing ratio, defined as the mass of water vapour divided by the mass of dry air. The extra term on the right hand side can be important if the water vapour flux is high, compared to the dry sensible heat flux. Use of the virtual temperature results in a sensible heat flux density which is greater than the dry sensible heat flux by 7% of the latent energy flux density LE. Appendix C shows that the additional thermodynamic energy of expansion should be 7.6% of the latent energy flux density, approximately independent of altitude. For example, a vigorously evaporating surface with an LE of 500 W m^{-2} would have an extra 'eddy expansion' term of 38 W m^{-2}, which would be accounted for within 1% (of 500 W m^{-2}) if one estimated the sensible heat flux using the sonic temperature (extra term of 35 W m^{-2}).

Other discussions concerning enthalpy energy exchange, and usage of a modified specific heat to account for the difference between moist air and dry air are given in Brook (1978), Reinking (1980), Webb et al. (1980), Businger (1982), Frank and Emmitt (1981), Sun et al. (1995), among others. This specific heat difference accounts for a smaller correction in magnitude, on the order of less than 1% (Webb et al., 1980), and therefore is not discussed further here. A different, more generalized approach using a thermodynamic energy balance integrated over a thin flat layer with non-zero volume by Sun et al. (1995) resulted in extra covariance terms (including some with pressure perturbations) which are difficult to evaluate in practice, so their method was not employed here.

One can approximate the enthalpy change and therefore the sensible heat flux exchange, in the differential form of enthalpy (Iribarne and Godson, 1981), instead of using the total integrated enthalpy, by the formal substitution of $C_p \delta T$ in place of mixing ratio s in Equation (9) through Equation (17), as suggested by Webb et al. (1980). Although the formal equations for the first law of thermodynamics generally involve the total derivative of temperature instead of the conservation equations for a scalar noted above, these two forms of equations are equivalent when the continuity equation is introduced, so the above substitution can be made,

$$\begin{aligned} s &= C_p \delta T, \\ s' &= C_p T'. \end{aligned} \tag{28}$$

Webb et al. (1980) note that

$$s' \approx \frac{\rho_c'}{\bar{\rho}} - \frac{\bar{\rho}_c}{\bar{\rho}^2}\rho', \tag{29}$$

which leads to the approximation,

$$\overline{w's'} \approx \frac{\overline{w'\rho_c'}}{\bar{\rho}} - \frac{\bar{\rho}_c}{\bar{\rho}^2}\overline{w'\rho'} \tag{30}$$

or for the temperature substitution,

$$\bar{\rho} C_p \overline{w'T'} \approx \overline{w'(\rho C_p T)'} - C_p \bar{T}\,\overline{w'\rho'}, \tag{31}$$

yielding the equivalent of Equation (19) for temperature,

$$\bar{\rho} C_p \frac{\partial \bar{T}}{\partial t} + \bar{\rho} C_p \bar{w} \frac{\partial \bar{T}}{\partial z} + \bar{\rho} C_p \frac{\partial \overline{w'T'}}{\partial z} = \bar{S}_T. \tag{32}$$

The horizontal advection gradient term can be included in Equation (32) if appropriate. Additionally, a block average of the moving average can be used for the first, transient term:

$$\bar{\rho} C_p \frac{\partial \overline{[T]}}{\partial t} + \bar{\rho} C_p \bar{w} \frac{\partial \bar{T}}{\partial z} + \bar{\rho} C_p \frac{\partial \overline{w'T'}}{\partial z} = \bar{S}_T . \tag{33}$$

8. Summary and Conclusions

We have presented here a theoretical basis for combining the Webb et al. (1980) correction with non-ideal advective effects and transient effects. Some simplified versions of the complete differential equations are made, with the classification of advection into two separate effects, that of a horizontal gradient of a scalar, and the other the horizontal gradient of the velocity field. Examples are given which show both the vertical advection term and the Webb et al. (1980) correction can be important for CO_2 exchange. An example is given showing that a reasonably flat, long fetch crop can experience a significant vertical advection component of sensible heat exchange. It is shown under transient conditions, conventional eddy covariance based on fixed-interval Reynolds averaging can be combined with other terms to provided a reasonable average of the exchange, even if a moving average is used for the time derivative of the scalar. In addition, a new correction is postulated for the energy budget, arising from the energy expended in expanding moist air under constant pressure conditions, which is automatically included in sensible heat eddy-covariance estimates using the sonic temperature.

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Appendix A

To perform a simple integration in the vertical of Equations (26) and (33), we assume that the values of s, ρ, T and the ratio of ρ and T are known at the bottom (soil surface, subscript 'o') and at the measurement height z_m (subscript 'm'), and that these variables change linearly between the two heights. The density covariance term becomes,

$$\int_o^{z_m} \bar{s}\frac{\partial(\overline{w'\rho'})}{\partial z}\mathrm{d}z \approx \int_o^{z_m} \bar{s}\frac{\bar{\rho}}{\bar{T}}\frac{\partial(\overline{w'T'})}{\partial z}\mathrm{d}z \approx \overline{w'T'}\left(\bar{s}_m\left(\frac{\overline{\rho_m}}{3\bar{T}_m}+\frac{\bar{\rho}_o}{6\bar{T}_o}\right) + \bar{s}_o\left(\frac{\bar{\rho}_m}{6\bar{T}_m}+\frac{\bar{\rho}_o}{3\bar{T}_o}\right)\right) \tag{A1}$$

and the advection term,

$$\int_o^{z_m} \bar{w}\bar{\rho}\frac{\partial(\bar{s})}{\partial z}\mathrm{d}z \approx \bar{w}_m(\bar{s}_m - \bar{s}_o)\left(\frac{\overline{\rho_m}}{3}+\frac{\bar{\rho}_o}{6}\right), \tag{A2}$$

where the subscript m indicates values at the measurement height, and o, atmospheric values at the soil surface.

Similar integrations can be performed if values of the profile are known at other heights, and with the horizontal mean scalar gradient terms; and the integrations can also be performed in the horizontal direction if appropriate.

Appendix B

The time-block average of the time derivative of s over an interval T, when the time derivative is taken with the forward-difference approximation, is

$$\overline{\frac{\partial s(t)}{\partial t}} = \frac{1}{n}\sum_{i=1}^{1=n}\frac{s(i+1)-s(i)}{\delta t} = \frac{s(n+1)-s(1)}{T}, \tag{B1}$$

where δt is the time interval between samples and n is the number of points in time interval T over which the block average is taken.

To determine whether the s term is important, one can carry out derivations similar to Equation (B1). The block average of the time derivative of the moving average of s is,

$$\frac{\overline{\partial[s(t)]}}{\partial t} = \frac{\bar{s}(n+1) - \bar{s}(1)}{T}, \tag{B2}$$

where $\bar{s}(n+1)$ is the block average of s evaluated with $(n+1)$ as the centre point (so the summation limits would be $(1+n/2)$ and $(1+3n/2)$), and $\bar{s}(1)$ is the block average of s evaluated with 1 as the centre point (so the summation limits would be $1 - n/2$ and $1 + n/2$).

From Equation (23),

$$\begin{aligned}\frac{\overline{\partial s(t)''}}{\partial t} &= \frac{\overline{\partial[s(t)]}}{\partial t} - \frac{\overline{\partial s(t)}}{\partial t} = \frac{\bar{s}(n+1) - \bar{s}(1)}{T} - \frac{s(n+1) - s(1)}{T} \\ &= \frac{s'(n+1) - s'(1)}{T},\end{aligned} \tag{B3}$$

where $s'(n+1)$ is the perturbation from the block average evaluated with $n+1$ as the centre point and $s'(1)$ is the perturbation from the block average evaluated with 1 as the centre point. If one assumes that s is a random variable, then the most likely values of $s'(n+1)$ and $s'(1)$ are zero, or in other words, if the ensemble average (indicated by triangular brackets), is now taken of Equation (B3), the result would be

$$\left\langle \frac{\overline{\partial s(t)''}}{\partial t} \right\rangle = \left\langle \frac{s'(n+1) - s'(1)}{T} \right\rangle \approx 0 \tag{B4}$$

so the most likely value of

$$\frac{\overline{\partial s(t)''}}{\partial t} \approx 0, \tag{B5}$$

which supports the notion that the last term on the right hand side of Equation (23) is negligible.

Appendix C

The Webb et al. (1980) correction implies energy associated with evaporation into the atmosphere, necessary for the expansion of eddy-parcels against an approximately constant pressure. One can estimate this energy starting with the thermodynamic equation for the energy of expansion given by

$$C_v \, \mathrm{d}T + p \, \mathrm{d}\alpha = \mathrm{d}Q, \tag{C1}$$

where C_v is the specific heat at constant volume, $\mathrm{d}T$ is a differential of temperature, $\mathrm{d}Q$ is the heat added to a mass, and $\mathrm{d}\alpha$ is a differential in specific volume. The second term on the left-hand-side represents the work of expansion. The substitution of eddy perturbations for differentials as done by Webb et al. (1980), yields, for the density of air change solely from the perturbation in water vapour:

$$p\alpha' = -p\frac{\mu\rho_v'}{\rho^2}, \tag{C2}$$

where the primes denote perturbations, ρ_v is the specific humidity, and μ is the ratio of the molecular mass of air to the molecular mass of water.

The thermodynamic equation adapted for perturbations, under adiabatic conditions, then yields the temperature perturbation T' equivalent to the energy needed for expansion,

$$T' \approx \frac{-p\mu p_v'}{C_v\bar{\rho}^2}. \tag{C3}$$

When this temperature perturbation is multiplied by the volumetric specific heat at constant pressure, one obtains the equivalent sensible heat needed for expansion,

$$H_e \approx \bar{\rho}C_p\frac{\bar{p}\mu\overline{w'\rho_v'}}{C_v\bar{\rho}^2} = \frac{C_p}{C_v}\frac{LE}{L}\frac{\bar{p}\mu}{\bar{\rho}} \approx 0.076LE. \tag{C4}$$

One interpretation of this derivation would be that the energy for expansion (or contraction) originates from fluctuations in air temperature in the eddies close to the site of the source or sink of water vapour, and the fluctuations are damped in the expansion/contraction process of the eddies/parcels in equilibration of pressure.

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