

Assessing the Eddy Covariance Technique for Evaluating Carbon Dioxide Exchange Rates of Ecosystems: Past, Present and Future

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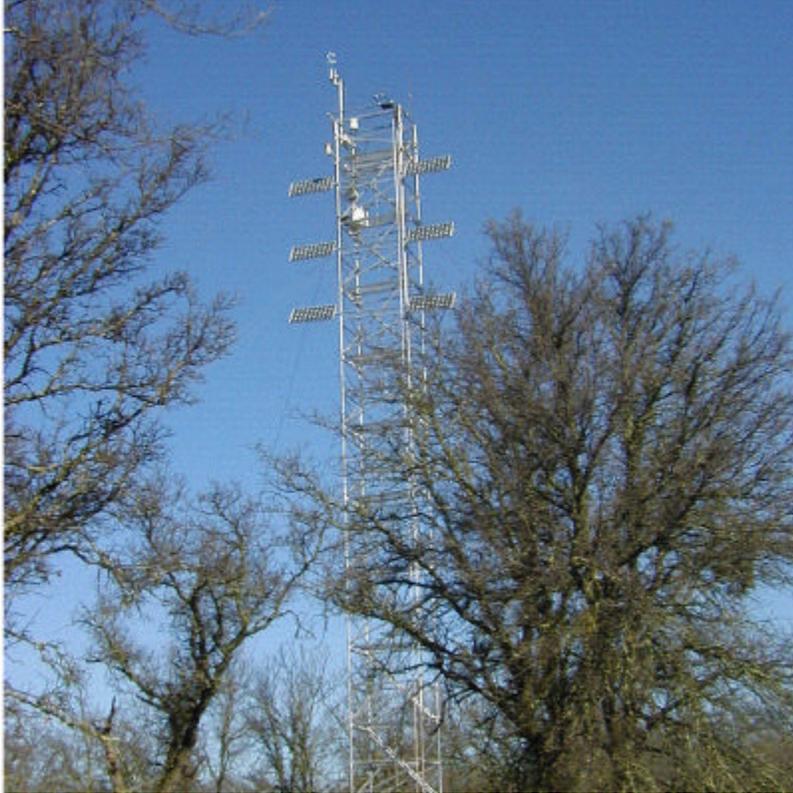
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Eddy Covariance Flux measurement system over an oak-grass savanna near

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ABSTRACT

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The eddy covariance technique ascertains the exchange rate of CO₂ across the interface between the atmosphere and a plant canopy by measuring the covariance between fluctuations in vertical wind velocity and CO₂ mixing ratio. Two decades ago, the method was employed to study CO₂ exchange of agricultural crops under ideal conditions during short field campaigns. During the past decade the eddy covariance method has emerged as an important tool for evaluating fluxes of carbon dioxide between terrestrial ecosystems and the atmosphere over the course of a year, and more. At present, the method is being applied in a nearly continuous mode to study carbon dioxide and water vapor exchange at over a hundred and eighty field sites, worldwide. The objective of this review is to assess the eddy covariance method as it is being applied by the global change community on increasingly longer time scales and over less than ideal surfaces.

The eddy covariance method is most accurate when the atmospheric conditions (wind, temperature, humidity, CO₂) are steady, the underlying vegetation is homogeneous and it is situated on flat terrain for an extended distance upwind. When the eddy covariance method is applied over natural and complex landscapes or during atmospheric conditions that vary with time, the quantification of CO₂ exchange between the biosphere and atmosphere must include measurements of atmospheric storage, flux divergence and advection.

Averaging CO₂ flux measurements over long periods (days to year) reduces random sampling error to relatively small values. Unfortunately, data gaps are inevitable when constructing long data records. Data gaps are generally filled with values produced from statistical and empirical models to produce daily and annual sums of CO₂ exchange. Filling data gaps with empirical estimates do not introduce significant bias errors because the empirical algorithms are derived from large statistical populations. On the other hand, flux measurement errors can be biased at night when winds are light and intermittent. Nighttime bias errors tend to produce an underestimate in the measurement of ecosystem respiration.

Despite the sources of errors associated with long-term eddy flux measurements, many investigators are producing defensible estimates of annual carbon exchange. When measurements come from nearly ideal sites the error bound on the net annual exchange of CO₂ is less than +/- 50 gC m⁻² year⁻¹. Additional confidence in long term measurements is growing because investigators are producing values of net ecosystem productivity that are converging with independent values produced by measuring changes in biomass and soil carbon, as long as the biomass inventory studies are conducted over multiple years.

INTRODUCTION

A prime focus of much research in the biogeosciences is on the net carbon balance of ecosystems (Running et al., 1999; Geider et al., 2001). Such work requires the assessment of carbon dioxide fluxes on hourly, daily, seasonal and yearly time scales and across the spatial scale of leaves, individual plants and arrays of plants.

Traditional tools used to assess net carbon exchange of ecosystem components include leaf cuvettes (Field et al., 1982; Collatz et al., 1991) and whole-plant (Denmead et al., 1993) and soil (Livingston and Hutchison, 1995; Goulden and Crill, 1997) chambers. The *forte* of cuvette and chamber systems is their ability to measure diurnal variations of carbon fluxes and to define environmental response functions (Schulze and Koch, 1969; Collatz et al., 1991). The physical placement of a cuvette on a leaf or a chamber over a plant or the soil, however, may produce biases and artifacts. For example, the manual dependency of a cuvette limits the number of leaves that can be measured across the domain of a plant canopy within a reasonable time frame, *e.g.* an hour. Hence, it is difficult to sample, with high statistical confidence, the natural variability that exists in photosynthesis; sources of variation include the acclimation of leaf photosynthesis to sun or shade environments and vertical gradients in photosynthetic capacity (Ellsworth and Reich, 1993). Similarly, the spatial extent that is sampled by a soil chamber, or a set of chambers, is relatively small compared to the spatial variability of the CO₂ efflux from the soil (Livingston and Hutchison, 1995; Law et al. 2001); the coefficient of spatial variation for soil respiration can reach 100% due to spatial gradients in soil texture, moisture, nutrients, temperature and roots.

Experimental artifacts introduced by cuvettes can be small if one is controlling the temperature, light, CO₂ and humidity, as is usually done when one is quantifying environmental response functions (Field et al., 1982; Collatz et al., 1991). On the other hand, placing a chamber over the soil to measure respiration by the rhizosphere introduces several bias errors. These include perturbations of local pressure, wind and CO₂ concentration fields and an alteration of the heat and water balance of the soil (Livingston and Hutchison, 1995; Lund et al., 1999; Davidson et al., 2002).

Enshrouding plants or trees in large transparent chambers (as is done when measuring the integrated carbon exchange of the plant-soil system) diffuses light, alters the canopy microclimate (Denmead et al., 1993), and suppresses soil respiration (Lund et al., 1999). Consequently, the shape of functions defining the response of canopy-scale, CO₂ exchange to environmental perturbations (as generated by whole-plant chambers) differ from those detected with independent micrometeorological measurements (Denmead et al., 1993; Ruimy et al., 1995).

The traditional means of addressing net ecosystem carbon exchange of an ecosystem over multiple years involves quantifying temporal changes of biomass (Clark et al., 2001) and soil carbon (Amundson et al., 1998; Lal et al., 2001). In principle, inventory studies of biomass change produce estimates of annual net primary productivity. Furthermore, forest biomass inventory studies rely on allometric relations to scale incremental changes in diameter at breast height to net primary production at plot and landscape scales (Clark et al., 2001; Barford et al., 2001). Bias errors are introduced when allometric relationships ignore trees in small size classes, understory vegetation, the amount of carbon that is allocated below ground (Clark et al., 2001) and when they do

not represent the multi-aged and multi-species structure of forest stands (Carey et al., 2001). There are also practical limitations with measuring temporal changes in soil carbon inventories. These arise from high degrees of spatial variability (vertical and horizontal) in bulk density and soil carbon (Lal et al., 2001). One also needs to quantify soil carbon in slow and fast pools (Amundson et al. 1998).

In recent years the eddy covariance technique has emerged as an alternative way to assess ecosystem carbon exchange (see Running et al., 1999; Canadell et al., 2000; Geider et al, 2001). Four factors account for this popularity. Most importantly, it is a scale-appropriate method because it provides ecosystem scientists with a method to assesses net CO₂ exchange of a whole ecosystem—one can consider it to be the canopy-scale equivalent to the cuvette, the primary tool for examining CO₂ exchange of leaves. Secondly, the eddy covariance technique produces a *direct* measure of net carbon dioxide exchange across the canopy-atmosphere interface. This task is accomplished by using micrometeorological theory to interpret measurements of the covariance between vertical wind velocity and scalar concentration fluctuations (Baldocchi et al., 1988; Verma, 1990; Desjardins, 1991; Lenschow, 1995). Thirdly, the area sampled with this technique, called the flux footprint, possesses longitudinal dimensions ranging between a hundred meters and several kilometers (Schmid, 1994). And finally, the technique is capable of measuring ecosystem CO₂ exchange across a spectrum of times scales, ranging from hours to years (Wofsy et al., 1993; Baldocchi et al., 2001a).

The eddy covariance method is particularly adept as studying ecosystem physiology. Specifically, it can be used to quantify how CO₂ exchange rates of whole ecosystems respond to environmental perturbations (Law et al., 2002), and when paired

systems are applied the method can be used to assess management questions such as the effects of disturbance and stand age (Chen et al., 2002; Anthoni et al., 2002) or plant functional type (Baldocchi and Vogel, 1996; Law et al., 2002).

Like cuvettes and chambers, the eddy covariance method has limitations, too. This method is most applicable: 1) over flat terrain; 2) when the environmental conditions are steady; and 3) when the underlying vegetation extends upwind for an extended distance. Violation of these assumptions can cause systematic errors in the interpretation of the eddy covariance measurements (Baldocchi et al., 1988; Foken and Wichura, 1995; Massman and Lee, 2002), which magnify when integrated over time to produce daily and annual sums (Moncrieff et al., 1996). In fact controversy has already occurred for this reason. The eddy covariance technique has attracted criticism from members of the ecological community because there have been cases when results produced from eddy covariance technique did not match estimates of net ecosystem productivity produced with established ecological methods (Keller et al., 1996; Piovesan and Adams, 2000). The eddy covariance research community is working feverishly to understand and remedy bias errors by carefully re-evaluating theory used to interpret flux measurements (Massman and Lee, 2002) and by conducting comparative studies with models (Wilson and Baldocchi, 2001) or biomass inventories to constrain flux estimates (Barford et al., 2001; Curtis et al., 2002).

Because so many advances have been made in the last decade and because many ecological questions require use of the eddy covariance method over extended time periods and over non-ideal landscapes there is a need to produce a contemporary review the eddy covariance technique and discuss its merits and limitations. To achieve this goal

I: 1) present an overview of the historical development of the technique; 2) discuss its theoretical foundation; 3) assess how it can be applied to construct daily and yearly sums of net carbon dioxide exchange between ecosystems and the atmosphere over ideal and non-ideal surfaces and 4) discuss directions of future research.

HISTORICAL DEVELOPMENT

Use of the eddy covariance technique has accelerated in recent years. A citation search of published papers that index the term ‘eddy covariance’ produced over 300 records and over 550 papers referred to the analogous and older term ‘eddy correlation’. The popular use of this method has not evolved spontaneously. Instead, it is built on a long history of fundamental research in the fields of fluid dynamics and micrometeorology and on technological developments associated with meteorological instruments, computers and data acquisition systems. In this section I give a brief survey of the history of research leading to contemporary application of the eddy covariance method.

Sir Osborne Reynolds is credited with establishing the theoretical framework for the eddy covariance technique (Reynolds, 1895). A lack of instrumentation, however, hindered the application of the eddy covariance method until 1926, when Scrase (1930) conducted a study on momentum transfer, the so-called Reynolds’ stress, with simple analog instruments and strip-chart data logging. The next wave of advancement in the eddy covariance technique came after World War Two, with the development of fast responding hot-wire anemometry and thermometry and digital computers (Swinbank, 1951). The first post-war eddy covariance studies were conducted over short vegetation at locales with extremely level terrain and windy, sunny climes and they focused on the

structure of turbulence in the atmospheric boundary layer and the transfer of heat and momentum, rather than on CO₂ exchange (Swinbank, 1951; Kaimal and Wyngaard, 1990). Nevertheless, these pioneering studies are notable for laying the theoretical and experimental foundation for subsequent work on measuring CO₂ exchange, which occurred during the late 1950's and early 1960's over short and ideal agricultural crops by Japanese, British and American scientists (Inoue, 1958; Monteith and Szeicz, 1960; Lemon, 1960). These first CO₂ exchange measurements, however, relied on the flux-gradient method (an indirect technique that evaluates flux densities of CO₂ as the product of a turbulent diffusivity (K) and the vertical gradient of CO₂ concentration, dc/dz), rather than the eddy covariance technique, due to a lacking of fast responding anemometers and CO₂ sensors.

The first CO₂ flux measurements made over forests (Denmead, 1969; Baumgartner, 1969; Jarvis et al., 1976) and native ecosystems, such as tundra, grasslands and wetlands (Coyne and Kelley, 1975; Ripley and Redman, 1976; Houghton and Woodwell, 1980), did not occur until the late 1960's and early 1970's. Application of flux-gradient theory over tall vegetation was found to be problematic at the onset (Raupach, 1979). Over tall forests vertical gradients of CO₂ are small and difficult to resolve because turbulent mixing is efficient. Secondly, evaluation of eddy exchange coefficients (K), using Monin-Obukhov similarity theory (Lenschow, 1995), is invalid above forests because turbulent transport is enhanced by the presences of a roughness sublayer (Raupach, 1979; Simpson et al., 1998). At this point in time, additional studies on CO₂ exchange over forests would need to wait for technical developments that would permit use of the eddy covariance technique.

The first eddy covariance measurements of carbon dioxide exchange occurred in the early 1970's (Desjardins, 1974; Desjardins and Lemon, 1974). This set of studies was performed over corn using a propeller anemometer and a modified, closed-path infrared gas analyzer, with a capacitance detector; a set of sensors with relatively slow time constants—on the order of 0.5 s. The slow time-response of the sensors used by Desjardins and Lemon (1974) prompted Garratt (1975) to critique those measurements and conclude that the CO₂ fluxes suffered from large errors (~40%) because they were unable to sample the high frequency portion of the flux cospectrum.

The next wave of technological improvements came nearly a decade later and relied on the commercial sonic anemometers and the development of rapid-responding, open path infrared gas analyzers (Jones et al., 1978; Bingham et al., 1978; Brach et al. 1981; Ohtaki and Matsui, 1982). Open-path CO₂ sensors, using solid-state, lead-selenium (PbSe) detectors, were a key innovation because they are able to sense CO₂ fluctuations as rapidly as ten times per second. And the open architecture of the CO₂ sensors enabled them to sample parcels of air with minimal aerodynamic disturbance. The first application of open-path CO₂ sensors was to measure eddy fluxes over crops. Among the first studies to apply this technology were those conducted by Anderson et al. (1984) over soybeans, Anderson and Verma (1986) over sorghum, Ohtaki (1984) over rice and Desjardins (1985) over corn. These initial efforts were soon followed by sets of experiments of CO₂ exchange over native vegetation, such as temperate deciduous forests (Wesely et al., 1983; Verma et al., 1986), a prairie grassland (Verma et al., 1989; Kim and Verma, 1990), a tropical forest (Fan et al., 1990) and Mediterranean macchia (Valentini et al., 1991).

Prior to 1990, limitations in sensor performance and data acquisition systems restricted the duration of the eddy covariance studies to short campaigns during the growing season (e.g. Anderson et al., 1984; Verma et al., 1986). Subsequent production of commercial infrared spectrometers, that were stable and had short time constants, enabled scientists to conduct eddy covariance measurements 24 hours a day, seven days a week, 52 weeks a year. Wofsy et al. (1993) are credited with conducting the first yearlong study of CO₂ exchange with the eddy covariance technique; these measurements were made over a deciduous forest, starting in 1990, and continue to this day. By 1993, a handful of additional eddy covariance studies, measuring CO₂ and water vapor exchange, began operating over forests in North America (Black et al., 1996; Goulden et al., 1996a, b; Greco and Baldocchi, 1996), Japan (Yamamoto et al., 1999), and Europe (Valentini et al., 1996). And by 1997, regional networks of flux measurement sites were operating in Europe (CarboEuroflux, Valentini et al., 2000; Aubinet et al., 2000) and North America (AmeriFlux, Running et al., 1999; Law et al., 2002). Currently, the eddy covariance method is being used at over 180 sites worldwide, as part of the FLUXNET program (Baldocchi et al, 2001) and involves new regional networks in North America (Fluxnet Canada), Brazil, Asia (AsiaFlux), Australia (OzFlux) and Africa.

PRINCIPLES AND CONCEPTS

The atmosphere contains turbulent motions of upward and downward moving air that transport trace gases such as CO₂. The eddy covariance technique samples these turbulent motions to determine the net difference of material moving across the canopy-atmosphere interface. In practice, this task is accomplished by statistical analysis of the instantaneous vertical mass flux density ($F = w\mathbf{r}_c$, $\mu\text{mol m}^{-2} \text{s}^{-1}$), using Reynolds' rules

of averaging (Reynolds, 1895), which are described below. The product of this operation is a relationship that expresses the mean flux density of CO₂ averaged over some time span (such as an hour) as the covariance between fluctuations in vertical velocity (w) and the CO₂ mixing ratio ($c = \frac{\mathbf{r}_c}{\mathbf{r}_a}$ where \mathbf{r}_a is air density and \mathbf{r}_c is CO₂ density):

$$F = \overline{\mathbf{r}_a \cdot w' c'} \quad (1)$$

In Eq. 1, the overbars denote time averaging and primes represent fluctuations from the mean (e.g. $c' = c - \bar{c}$). A positively signed covariance represents net CO₂ transfer into the atmosphere and a negative value denotes the reverse.

Interpreting Eddy Covariance Measurements

The equation defining the conservation of mass provides theoretical guidance for implementing the eddy covariance technique (Baldocchi et al., 1988; Paw U et al., 2000; Massman and Lee, 2002). Conceptually, the problem being analyzed is analogous to the case of maintaining a certain water level in a bathtub by governing the flow of water in and out of the tub; when the water level is steady, we know that the rate of water leaving the tub equals the rate entering so we only need to measure one of the flows. With the case of assessing turbulent transfer of CO₂ in the atmosphere, we use the conservation equation to deduce the exchange of carbon in and out of the plant-soil system on the basis of eddy covariance measurements made in the surface boundary layer several meters above a plant canopy. How this process is accomplished is discussed next.

For the case of CO₂, the conservation of mass states that the sum of the local time rate of change of the CO₂ mixing ratio (term I), c , and advection (term II) is balanced by

the sum of the flux divergence of CO₂ in the vertical (z), lateral (y) and longitudinal (x) directions (term III) and the biological source sink-strength (S_B) (term IV):

$$\begin{aligned} \frac{d\bar{c}}{dt} &= \frac{\partial \bar{c}}{\partial t} + \underbrace{u \frac{\partial \bar{c}}{\partial x}}_I + \underbrace{v \frac{\partial \bar{c}}{\partial y}}_II + \underbrace{w \frac{\partial \bar{c}}{\partial z}}_III \\ &= - \left(\underbrace{\frac{\partial F_z}{\partial z}}_III + \underbrace{\frac{\partial F_x}{\partial x} + \frac{\partial F_y}{\partial y}}_IV + S_B(x, y, z) \right) \end{aligned} \quad (2)$$

In Equation 2, *u*, *v* and *w* are the vertical velocities in the *x*, *y* and *z* directions, respectively. Figure 1 is a conceptual diagram of the processes described by Equation 2.

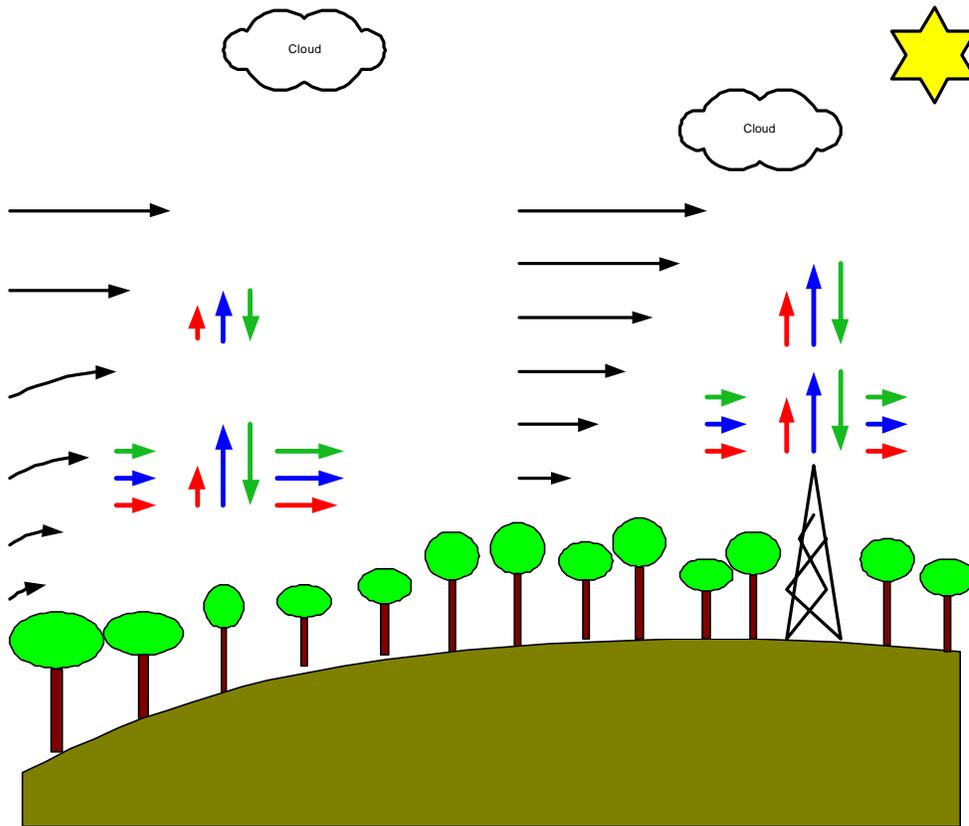


Figure 1 Conceptual diagram of an eddy flux tower on terrain that experiences a transition from less ideal sloping to ideal flat region. Flux densities of water (blue arrows), heat (red) and CO₂ (green) are generally orthogonal to the wind streamlines (black arrows). On the upwind side of the hill wind accelerates as the hill compresses the wind velocity streamlines. Advection occurs and produces divergence of the fluxes of heat, water vapor and CO₂, causing the vertical exchange to vary with height (denoted by

changes in arrow length). Over the flat part of the landscape, a constant flux layer and logarithmic wind profile are re-established. Fluxes of mass and energy are constant with height as denoted by arrows of similar length at two different heights above the canopy.

Under ideal conditions, the scalar concentrations and wind velocities in the atmosphere are steady with time (term I equals zero) and the underlying surface is horizontally homogeneous and on flat terrain (there is no advection, term II) so the horizontal flux divergences, $\frac{\partial F_x}{\partial x}$ and $\frac{\partial F_y}{\partial y}$, in term III equal zero). Based on these assumptions, the conservation equation simplifies to a balance between the vertical flux divergence of CO₂ and its biological source/sink strength, S_B , (term IV):

$$\frac{dF_z}{dz} = -S_B(z) \quad (3)$$

By integrating Equation 3 with respect to height, one derives an equality between the mean vertical flux density measured at some height above the canopy, $F_z(h)$, and the net flux density of material in and out of the underlying soil, $F_z(0)$, and vegetation:

$$F_z(h) = F_z(0) - \int_0^h S_B(z) dz \quad (4)$$

In practice it is the term $F_z(h)$ that is evaluated with the eddy covariance technique.

Evaluating the Flux Covariance

Assessment of the flux covariance requires that we sample the cospectrum of turbulent motions that exist in the atmosphere (Garratt, 1975):

$$\overline{w'c'} = \int_0^{\infty} S_{wc}(\mathbf{w}) d\mathbf{w} \quad (5)$$

where S_{wc} is cospectral density between w and c (the amount of flux that is associated with a give frequency) and ω is angular frequency, (ω is related to natural frequency, n , by a factor of 2π). To sample all significant scales of motion that contribute to turbulent transfer of CO_2 , one needs to sample the atmosphere frequently and for a sufficient duration. Sampling rates on the order of 10 times per second generally enable one to sample the high frequency portion of the flux co-spectrum (Anderson et al., 1984, 1986; Goulden et al., 1996a). To capture low frequency contribution to the flux covariance, the sampling duration must be long enough to sample motions associated with the convective boundary layer, but the sampling duration should not be too long to be affected by diurnal changes in CO_2 (Lenschow, 1995). It is common practice to sample and average atmospheric turbulence over 30 to 60 minutes periods during daylight hours (Anderson et al., 1984, 1986; Aubinet et al., 2000; Massman and Lee, 2002). Longer averaging times may be needed at night when the thermal stratification of the atmosphere is stable and turbulence is intermittent (Lee et al. 1996; Massman and Lee, 2002).

In practice, numerous instrument, sampling and turbulence issues influence how well the co-spectrum integral (Eq 5) is measured. The w - c covariance, that is measured by a set of instruments, is a function of its true cospectral density, $S_{wc}(\mathbf{w})$, and a filter function, $H(\mathbf{w})$ (Moore, 1986; Leuning and Judd, 1996; Aubinet et al., 2000; Massman, 2000; Berger et al., 2001):

$$\overline{w'c'}_{measured} = \int_0^{\infty} H(\mathbf{w})S_{wc}(\mathbf{w})d\mathbf{w} \quad (6)$$

High pass filtering of the turbulence signals (w or c) (the attenuation of high frequency contributions to the flux) can be caused by a sensor's slow response time, a long sensor

path, or a slow sampling rate. High pass filtering will also occur if wind velocity and trace gas sensors are placed too far abreast from one another or too close to the ground (Moore, 1986; Massman, 2000; Aubinet et al., 2000). A third means of filtering high frequency scalar fluctuations involves sampling air through a tube (Sukyer and Verma, 1993; Leuning and Judd, 1996; Massman, 2000).

Low pass filtering (the attenuation of low frequency contributions to the flux) is imposed by the averaging method used to compute the flux covariance and the sampling duration (Moore, 1986; Massman, 2000; Aubinet et al., 2000). Low pass filtering is most severe during convective conditions, when the passage of a large convective cell may be incomplete during a conventional averaging period.

How one computes means and fluctuations from the mean is an issue that affects the computation of Eq 1. It is common practice to fluctuations by removing the arithmetic mean or one determined with a digital recursive filter (McMillen, 1988) from instantaneous measurements. Some investigators detrend signals (Foken and Wichura, 1995) before computing fluctuations from the mean. This author is not an advocate of detrending turbulent signals because he considers it to be redundant to detrend turbulence signals and to compensate for storage of CO₂ in the canopy air space (see below). Relying on fundamental grounds, Reynolds' averaging rules do not consider detrending when defining flux covariances (Reynolds, 1895; Paw U et al., 2000). The reader should recognize that detrending is a historical artifact that needs careful evaluation before applying. When the eddy covariance method was being developed it was necessary to detrend electrical signals because instruments suffered from electronic drift

(Shuttleworth, 1988). Sensors are now stable, so this need has been eliminated for the most part.

At present, options for making accurate eddy covariance measurements are either to design an eddy covariance system that minimizes cospectral filtering or to assess the spectral transfer function (Eq 6) and correct one's measurements. Moore (1986) and Massman (2000) have produced theoretical transfer functions that can be applied to correct eddy covariance measurements. In general, spectral corrections factors range between 1.04 and 1.25 for CO₂ flux densities measured by systems employed by the Euroflux community (they use closed path gas analyzers) and the spectral corrections range between 1.06 and 1.35 for water vapor flux densities (Aubinet et al., 2000). Alternatively, Wofsy et al. (1993) and Goulden et al. (1996a, b) use an empirical approach to correct eddy covariance measurements. They correct the measured flux covariance by the ratio of a nearly 'perfect' measure of covariance, such as that between acoustic temperature and w , and its artificially 'degraded' value. The time constant of a low pass recursive filter is adjusted to degrade the 'perfect' signal until it mimics the measured CO₂ signal. This method does not rely on any assumptions about the functions describing turbulence spectra and transfer functions. On the other hand, it does not account for line averaging across the fixed distance of the anemometer path and it fails when sensible heat flux density is near zero.

The analytical method used to measure CO₂ also has an impact on the computation of the flux covariance. In practice, CO₂ is measured with a non-dispersive, infrared spectrometer. This sensor does not measure mixing ratio, c . Instead it samples molar density, r_c (moles per unit volume). In principle, changes in molar density can

occur by adding molecules to or removing them from a controlled volume or by changing the size of the controlled volume, as is done when pressure, temperature and humidity change in the atmosphere. By measuring the eddy flux covariance in terms of molar density, the net flux density of CO₂ across the atmosphere-biosphere interface is re-expressed as:

$$F = \overline{w\mathbf{r}_c} = \overline{w'\mathbf{r}_c'} + \overline{w\mathbf{r}_c} \quad (7)$$

I II

The new term, on the right hand side of Eq. 7 (term II), is the product of the mean vertical velocity and CO₂ density. The mean vertical velocity is non-zero and arises from air density fluctuations (Webb et al., 1980; Kramm et al., 1995). In practice, the magnitude of \overline{w} is too small ($< 1 \text{ mm s}^{-1}$) to be detected by anemometry, so it is usually computed on the basis of temperature (T) and humidity density (\mathbf{r}_v) fluctuations using the Webb-Pearman-Leuning (1980) algorithm:

$$F_c = \overline{w'\mathbf{r}_c'} + \frac{m_a}{m_v} \frac{\overline{\mathbf{r}_c}}{\overline{\mathbf{r}_a}} \overline{w'\mathbf{r}_v'} + \left(1 + \frac{\overline{\mathbf{r}_v m_a}}{\overline{\mathbf{r}_a m_v}}\right) \frac{\overline{\mathbf{r}_c}}{\overline{T}} \overline{w'T'} \quad (8)$$

(other variables in Equation 8 are the molecular weights of air, m_a , and water vapor, m_v)

The derivation of Equation 8 ignores effects of pressure fluctuations, which may be significant under high winds (Massman and Lee, 2002), and covariances between temperature and pressure (Fuehrer and Friehe, 2002). It also ignores advection (Paw U et al., 2000), which will be important when it is applied over sloping terrain. Despite the assumptions used in deriving Eq. 8, there is experimental evidence supporting its validity. Leuning et al. (1982) measured CO₂ exchange over a flat, bare, dry field and found that photosynthesis was ‘detected’ when they did not apply the density corrections. In

contrast, application of Eq. 8 produced CO₂ flux densities near zero, a value supported with independent chamber measurements.

Significant terms in Equation 8 depend on whether one uses an open or closed path infrared spectrometer. If one draws air down a heated tube in a turbulent state, as is needed to implement a closed-path sensor, temperature fluctuations will dampen and approach zero, thereby canceling the last term on the right hand side of Equation 8 (Leuning and Moncrieff, 1990; Leuning and Judd, 1996).

Evaluating the Biosphere-Atmosphere Exchange Rates

There are times during the day when the flux density of CO₂ crossing a horizontal plane above the canopy does not equal the net flux density of carbon moving into and out of the plant/soil system. At night, for example, the thermal stratification of the atmosphere is stable CO₂. Under this condition, CO₂ exiting leaves and the soil may not reach a set of instruments at a reference height, h , above the canopy, causing the eddy covariance method to underestimate ecosystem respiration. Another case occurs at sunrise. Then, there is a break-up of the stable nocturnal boundary layer as convective turbulence resumes. This phenomenon vents the canopy of CO₂ stored within the canopy air space over the course of the night. This short-term venting will cause the eddy covariance technique to overestimate the time-local flux density (Grace et al., 1996; Moncrieff et al., 1996; Yang et al., 1999). Under the non-steady conditions identified

above, the storage of CO₂ in the underlying airspace, $\overline{r_a} \int_0^h \frac{\partial \bar{c}}{\partial t} dt$, is non-zero. Storage must be assessed and added to the eddy covariance measurement if we expect to obtain a measure of the net flux of CO₂ flowing into and out of the soil and vegetation (Fan et al., 1990; Grace et al., 1996; Goulden et al., 1996; Moncrieff et al., 1996; Yang et al., 1999;

Baldocchi et al., 2000; Aubinet et al., 2000). To measure the storage term accurately, one must measure temporal changes in CO₂ above the canopy and, at least, two heights within the canopy (Yang et al., 1999). On daily and annual time scales the storage term is approximately zero so errors in its evaluation are not critical (Anthoni et al., 1999; Baldocchi et al., 2000).

Over sloping terrain, the mean vertical velocity, with respect to the geopotential, will be non-zero because hills cause wind streamlines to converge and diverge (Finnigan, 1999; Figure 1). Classical application of the eddy covariance method involves mathematical rotation of the wind coordinate system to force \overline{w} to zero. This rotation enables one to compute flux covariances that are orthogonal to the mean streamlines flowing over the landscape (Wesely, 1970; Baldocchi et al., 1988; Foken and Wichura, 1995). If mesoscale circulations persist, there are wind biases introduced by the eddies shed from the instrument tower or zero-offsets associated with the anemometer, it is inappropriate to rotate the coordinate system and force the mean vertical velocity to zero (Lee, 1998). A new reference for coordinate rotation must be defined which will depend on wind direction, instrument biases and the slope of the upwind terrain (Paw U et al., 2000).

Wind flow over non-uniform terrain can also generate advective fluxes (Figure 1). From a practical standpoint, it is difficult to assess horizontal advection terms over tall forests. Motivated by this problem, Lee (1998) developed simplified version of the conservation of mass equation (Eq 2) that represents net ecosystem-atmosphere CO₂ exchange (N_e) as a function of a one-dimensional, vertical advection term. General application of the model of Lee (1998) is still subject to debate. Theoretically, Finnigan

(1999) argues that topography and spatial changes in surface roughness and CO₂ sources and sinks strengths, produce spatial variations in the scalar concentration and wind velocity fields that may not be accommodated by this one-dimensional advection equation. From a practical standpoint, the introduction of an additional term to the conservation budget introduces new sources of measurement error (Baldocchi et al., 2000). Alternative approaches to assessing CO₂ advection include measuring flux divergence profiles over and under plant canopies (Lee et al. 1999; Baldocchi et al. 2000, Yi et al. 2000), measuring horizontal transects across landscapes (Baldocchi and Rao, 1995) or by assessing regional box budgets (Sun et al., 1998; Eugster and Siegrist, 2000). As a word of caution, drainage flows may transport CO₂ from the vicinity of the eddy covariance measurement tower and vent it elsewhere (Sun et al., 1998), thereby leading to a systematic bias error and an underestimate of ecosystem respiration.

EVALUATING DAILY AND ANNUAL CARBON FLUXES

The Influence of Random and Systematic Bias Errors

Because of our desire to sum eddy fluxes over very long durations, as is needed to address ecologically relevant questions, we face new challenges to the application of the eddy covariance technique. In this section, we assess potential errors and discuss whether they are acceptable and if they cancel over longer integration times.

In practice, the accuracy of summing short-term eddy flux measurements on daily, seasonal and annual time scales depends upon a set of random and systematic bias errors that are associated with measurements, sampling and theoretical issues relating to the application of the eddy covariance technique to non-ideal conditions. With proper system design and implementation, random measurement errors are generally small.

For example, calibration errors of infrared gas analyzers are on the order of 2 to 3%. And errors associated with time lags between velocity and scalar sensors are less than 2% (Berger et al. 2001). In general, the covariance measurement error is less than 7% during the day and less than 12% at night (Moore, 1986; Berger et al, 2001; Soegaard et al., 2000).

The natural variability of turbulence is on the order of 10 to 20% (Wesely and Hart, 1985) and sets a limit on the run-to-run variability of flux measurements under similar conditions. Averaging numerous flux density measurements to construct longer-term averages (e.g. daily, weekly, monthly) reduces random sampling errors to a value within +/- 5%, thereby increasing the precision of CO₂ flux measurements (Moncrieff et al., 1996; Goulden et al. 1996a).

There is a practical limit with the concept of constructing long-term averages. Gaps in long-term data records will inevitably occur as sensors break down, they are being calibrated or when measurements over range the data acquisition system. Furthermore, data are generally rejected when the wind is blowing through a tower, when wind is coming from an undesirable wind sector, when sensors are wet, or when the measurements fail to meet pre-set acceptance criteria (Foken and Wichura, 1995). Typical data coverage, over the course of a year, ranges between 65 % and 75% for a large number of field studies (Falge et al., 2001); data gaps tend be lower for systems that employ closed path CO₂ sensors because they do not need to reject data because the sensor is wet.

Several methods are being employed to fill data gaps. One approach fills missing flux data on the basis of empirically derived algorithms that are driven by easily

measured meteorological variables, such as sunlight, temperature and humidity (Goulden et al., 1996b; Aubinet et al., 2000; Falge et al., 2001). This approach, however, needs continual updating and tuning because seasonal changes in leaf area, soil moisture, and photosynthetic capacity will alter any empirical relation. Another approach involves interpolation between adjacent periods. This method may work well for small data gaps but it will have problems with gaps occurring over several hours and days. A third approach is to bin data by hour for a one to two week period, then use the time-dependent mean to replace missing data (Moncrieff et al., 1996; Jarvis et al., 1997). Falge et al. (2001) compared several gap filling methods and found that they produced similar results and did not introduce any particular methodological bias. Falge et al. (2001) also found that rejecting up to 40% of data produces repeatable annual sums, as there is an adequate sampling of the data population.

Two types of systematic bias errors continue to plague eddy covariance measurements. One is the widely observed lack of energy balance closure (Aubinet et al., 2000; Twine et al., 2000; Wilson et al., 2002); practitioners often find that the sum of latent and sensible heat exchange, measured with the eddy covariance technique do not match the independent measurement of available energy. The other systematic bias error is associated with a perceived underestimate of nocturnal ecosystem efflux during low wind conditions (Grace et al., 1996; Black et al., 1996; Goulden et al., 1996a; Moncrieff et al., 1996; Mahli et al., 1998; Aubinet et al., 2000; Baldocchi et al., 2000). Tests of surface energy balance closure suggest that turbulent fluxes at some sites are systematically 10% to 30% too small to close the energy budget, (Aubinet et al., 1999; Twine et al., 2000; Wilson et al., 2002). These results raise the possibility that CO₂ fluxes

are underestimated, too. For investigators using open-path CO₂ sensors, errors in energy balance closure will translate into an additional source of errors in assessment of the Webb et al (1980) density corrections (Eq. 8), which are a function of sensible and latent heat flux densities. Factors contributing to a lack of energy balance closure include: 1) filtering of low frequency flux contributions; 2) advection; and 3) different footprints viewed by the eddy flux and the available energy measurement systems (Twine et al., 2000; Yi et al., 2000; Wilson et al., 2002).

Some researchers advocate adjusting CO₂ flux densities in proportion to the lack of energy balance closure (Twine et al. 2000). But this procedure places high levels of confidence on the accuracy and representativeness of the measurement of available energy; net radiometers and soil heat flux plates sample a small portion of the landscape near the tower, while eddy covariance measurements represent an area hundreds of meters square in area (Schmid, 1994). In addition, independent tests of evaporative fluxes, based on lysimeters and watershed water balances, agree well with eddy covariance measurements (Barr et al., 2000; Wilson et al., 2001), which lend support to the accuracy of daytime eddy covariance measurements.

Insufficient turbulent mixing, incorrect measurement of the storage term of CO₂ in the air space and soil, and the drainage of CO₂ out of the canopy volume at night have been posited as reasons why the eddy covariance method underestimates CO₂ flux densities at night (Black et al., 1996; Lindroth et al., 1998; Sun et al., 1998). At present, it is common practice to apply an empirical correction to compensate for the underestimate of nighttime carbon flux measurements. Some investigators replace data with a temperature-dependent respiration function that is derived from soil chambers

(Anthoni et al., 1999). Others correct nocturnal CO₂ flux density measurements with values measured during windy periods using a regression between CO₂ flux density and friction velocity (Black et al., 1996; Goulden et al., 1996a, b; Lindroth et al., 1998; Hollinger et al., 1999; Malhi et al., 1998; Aubinet et al., 2000; Falge et al., 2001; Lafleur et al., 2001). The critical friction velocity that produces ‘good’ nighttime CO₂ fluxes is not universal and can range from 0.1 to 0.6 m s⁻¹.

There are pro and cons to adjusting nocturnal CO₂ fluxes relative to values measured during well-mixed conditions. On the negative side, the regression statistics between F_c and friction velocity, u_* , tend to be poor; typically, the coefficient of determination (r^2) is less than 0.2 (Aubinet et al., 2000). Longer averaging periods reduce the sampling error of individual data points (Sukyer and Verma, 2001), but uncertainty in the statistical regression between F_c and the independent variable, friction velocity, remains high because of the reduced number of samples (Aubinet et al., 2000, 2001).

Despite the problems cited above, several investigators have reported favorable comparisons between nocturnal respiration rates (averaged over long periods) and independent estimates, as inferred by the intercept of the response curve between CO₂ flux and sunlight (Hollinger et al., 1999; Lee et al., 1999; Baldocchi et al., 2001; Sukyer and Verma, 2001; Falge et al., 2002). The zero intercept of the light response curve is a stable estimate of ecosystem respiration because it is determined using daytime flux measurements when mixing is better and the flux measurements are more accurate. And, respiratory fluxes averaged over the whole night tend to capture more eddies that transfer CO₂, as opposed to individually hourly runs that experience much variability.

Comparing nocturnal eddy flux measurements with scaled estimates of ecosystem respiration components using soil, bole and plant chambers, on the other hand, have produced mixed results. Lavigne et al. (1997), studying an array of boreal forest stands (10 to 15 m tall), and Law et al (1999), studying a 34 m ponderosa pine forest, report poor agreement between nocturnal eddy covariance measurements and scaled respiration measurements. The assessment of these studies, however, needs qualification. Lavigne et al. (1994), for example, relied on a relatively small sample size of soil respiration measurements and that study was conducted during the early years of long-term eddy covariance measurements (circa 1994). The independent of assessment of nighttime term eddy fluxes has improved since that time. A more recent comparative study produced good agreement between the nocturnal eddy covariance and chamber-based measurements of nocturnal respiration (Law et al., 2001). Ultimately, the validity of nocturnal flux measurements will depend on whether or not they produce ecologically defensible ratios between net carbon gain and losses (Law et al., 1999; Falge et al., 2001) and if they produce Q_{10} respiration coefficients that are consistent with those determined with other techniques.

Error Bounds on Annual Carbon Flux Sums

Several groups have produced formal error analyses on annual sum of net ecosystem CO_2 exchange. Goulden et al. (1996a) concluded that the sampling error, with 90% certainty, was $\pm 30 \text{ gC m}^{-2} \text{ y}^{-1}$ at Harvard forest, a site with an net annual uptake of CO_2 that is on the order of $200 \text{ gC m}^{-2} \text{ y}^{-1}$. Similar error sums have been computed by Lee et al. (1999) for a temperate broadleaved forest ($\pm 40 \text{ gC m}^{-2} \text{ y}^{-1}$), by Yang et al. (1999) for a boreal aspen stand ($\pm 30 \text{ gC m}^{-2} \text{ y}^{-1}$) and by Lafleur et al. (2001) for a short

bog ($\pm 68 \text{ gC m}^{-2} \text{ y}^{-1}$); all studies were on relatively on level terrain. Larger errors have been reported over less ideal sites. At Walker Branch Watershed, where I have conducted much of my research, the topography is hilly and the site is near two power plants. We have bounded the annual sum of CO_2 exchange using soil chambers, biophysical model calculations and advection estimates. This procedure produced a range of carbon fluxes of $\pm 130 \text{ gC m}^{-2} \text{ y}^{-1}$ for a site that takes up about $600 \text{ gC m}^{-2} \text{ y}^{-1}$ (Baldocchi et al., 2000; Wilson and Baldocchi, 2001). And a larger error bound has been reported for a ponderosa pine stand in even rougher terrain (Anthoni et al., 1999). This site was positioned along a pronounced ridge and has a bounded error term of its annual net carbon exchange of $\pm 180 \text{ gC m}^{-2} \text{ y}^{-1}$ on its annual net carbon uptake of about $300 \text{ gC m}^{-2} \text{ y}^{-1}$.

Comparisons between annual estimates of net ecosystem carbon exchange using eddy covariance measurements and traditional ecological methods are another way to assess bias errors. At this moment, only a few comparisons have been produced and they have generated mixed results (Curtis et al, 2002). On the favorable side, eddy covariance measurements by Schmid et al. (2000) agree within 5% of values determined using traditional ecological methods. And reports by Barford et al (2001), Ehman et al., (2002) and Curtis et al. (2002) indicate that annual sums of CO_2 produced by the eddy covariance method agree within 30% of biomass studies. A less successful study was reported by Granier et al (2000), who found a discrepancy of about 100% between net ecosystem productivity (NEP) measured with the eddy covariance method and annual biomass increment of stand (a difference of about $200 \text{ gC m}^{-2} \text{ year}^{-1}$); their carbon balance however did not consider respiration of woody debris. The emerging trend

being generated by contemporary field studies is a convergence between eddy covariance and biomass inventories when data records are evaluated over multiple years (Barford et al., 2001; Ehman et al., 2002; Chen et al., 2002).

FUTURE PROSPECTS

Much progress has been made in applying and interpreting eddy covariance measurements for ecological problems over the past two decades. At present, annual carbon budgets produced by eddy covariance measurements are most trustworthy when they come from micrometeorologically ideal sites, extensive canopies on flat terrain.

Eddy covariance measurements made over non-ideal sites have value, too, even though annual estimates of net CO₂ exchange may be error prone. Flux measurements from complex sites can provide information on the relationship between carbon fluxes and phenology, they can quantify how stand-scale carbon fluxes respond to environmental perturbations and they can quantify the factors causing year-to-year variability in carbon fluxes (e.g. Baldocchi et al. 2001b). At non-ideal sites, a better understanding of bias errors due to advection and drainage will be needed to reduce uncertainties in annual carbon fluxes from between 100 to 200 gC m⁻² year⁻¹ to less than 50 gC m⁻² year⁻¹. Future studies of CO₂ exchange on complex terrain should employ a combination of approaches (mesoscale modeling, remote sensing, ecophysiological measurements) to constrain fluxes. One example would involve using a network of sapflow measurements of trees on sloping terrain and scale these measurements to photosynthesis using estimates of water use efficiency that is derived from cuvettes or isotope content of leaves. Another example would involve explicit measurements of advection terms, storage and drainage terms.

Another priority for eddy covariance research will involve producing annual sums of carbon exchanges that represent scenes viewed by satellites (Running et al., 1999). To evaluate carbon fluxes over mosaics of vegetation we will need to quantify how carbon dioxide fluxes, the underlying vegetation and the flux footprint vary with wind direction (Amiro, 1998; Schmid and Lloyd, 1999; Soegaard et al. 2000; Aubinet et al. 2001). To accomplish this task, we will need to partition net carbon fluxes into the components, GPP, NPP and ecosystem respiration, so field data generated at eddy covariance study sites can be used to validate products developed by biogeochemical models and remote sensing indices (Running et al., 1999). To perform landscape integration of carbon fluxes we also need to understand how they vary with stand age, a problem that will require more chronosequence studies. Ultimately, confidence in long-term carbon flux measurements will come by using multiple constraints to interpret the annual sums. To meet this end, more collaborative studies among scientists working with the eddy covariance method, models, chambers and soil and biomass inventories will be needed.

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