Trace Gas Exchange Above the Floor of a Deciduous Forest 1. Evaporation and CO₂ Efflux

DENNIS D. BALDOCCHI AND TILDEN P. MEYERS

NOAA Atmospheric Turbulence and Diffusion Division, Oak Ridge, Tennessee

The eddy correlation method has great potential for directly measuring trace gas fluxes at the floor of a forest canopy, but a thorough validation study has not been yet conducted. Another appeal of the eddy correlation method is its ability to study processes that regulate and modulate gas exchange between the soil/litter complex and the atmosphere that cannot be probed with chambers. In this paper we report on eddy correlation measurements of water vapor, sensible heat, and carbon dioxide exchange that were made at the floor of a deciduous forest. The validity of the eddy correlation method to measure the emission of water vapor and CO₂ from a deciduous forest floor is demonstrated by our ability to close the surface energy budget during periods that meet the requirements of the technique. Water vapor fluxes from a dry forest floor are strongly influenced by large-scale turbulent events that penetrate deep into the canopy. The frequency of these turbulent events prevents equilibrium evaporation rates from being achieved because the dynamic time constant for water vapor exchange is longer. Consequently, maximal evaporation rates are capped to rates defined by the product of the driving potential of the atmosphere and the surface conductance. On the other hand, evaporation from a wet forest floor proceeds at rates reaching or exceeding equilibrium evaporation and are highly correlated with static pressure fluctuations. CO₂ efflux rates are governed by litter and soil temperature, as expected. But we also find a significant correlation between static pressure fluctuations and soil/litter CO2 exchange rates.

INTRODUCTION

Soil and litter are significant sources of water vapor and CO₂ in terrestrial ecosystems [Denmead, 1984; Walker, 1984; Black and Kelliher, 1989; Schlesinger, 1977]. The evaporation from the soil-litter complex is influenced primarily by available energy, soil moisture, atmospheric turbulence and humidity, and thermal and hydraulic soil properties [Kimball, 1983; Denmead, 1984; Walker, 1984; Black and Kelliher, 1989]. Biological processes, such as root respiration and decomposition of organic matter by microbial communities, produce the CO₂ that is released from the soil and litter. The amount and nutrient quality of substrate material are biotic variables that regulate the rates of microbial CO₂ production [Meentemeyer, 1978; Larkin and Kelly, 1987; Paul and Clark, 1989]. In turn, the quality and amount of litter is influenced by site fertility [Schlesinger, 1977; Vitousek, 1984]. Abiotic variables regulate CO₂ exchange by affecting the diffusion and mass transfer of gaseous compounds through the soil and litter and by influencing the rates of microbial and root respiratory activity. Important abiotic variables include the temperature and moisture and oxygen content of the soil and litter, atmospheric pressure fluctuations, and thermal and hydraulic soil properties [Galbally, 1989; Kimball, 1983; Schlesinger, 1977; Meentemeyer, 1978].

The rates of soil-atmosphere gas exchange from bare soils differ from soils covered with litter and situated under vegetated canopies. Litter acts as a mulch and a substrate for microbial decomposition. Mulches impede the diffusion of water vapor and other trace gases from the soil [Denmead, 1984; Rosenberg et al., 1983]. Litter mulches also insulate the soil and thereby alter the energy balance and temperature and moisture regimes [Rosenberg et al., 1983]. The

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presence of a vegetation canopy alters the radiation, wind, temperature, precipitation, and humidity regimes that drive soil evaporation and the biogenic emission of CO_2 from the soil/litter complex.

Biogeochemical cycling studies on the biogenic emission of carbon, nitrogen, and sulfur compounds from the soil system under plant canopies rely heavily on static and dynamic chamber methods [Garratt and Cox, 1973; Edwards et al., 1989; Williams et al., 1987; Goldan et al., 1988]. Chamber methods can introduce errors and biases on the measurement of soil-atmosphere trace gas emissions unless extreme care is taken. The most common errors relate to artifacts induced by static and dynamic chamber systems, disturbance of the local microenvironment, and inadequate spatial sampling due to large spatial variability [Mosier, 1989; Edwards and Sollins, 1973; Kanemasu et al., 1974; Cropper et al., 1985].

Most studies of soil evaporation below plant canopies have been based on microlysimeter technology [Walker, 1984; Black and Kelliher, 1989; Kelliher et al., 1990; Allen, 1990]. Lysimeter measurements are accurate if care is made not to adversely affect the temperature and moisture regimes and the bulk properties of the soil [see Tanner, 1967]. However, the method suffers from problems of adequate spatial sampling and representativeness due to hydraulic isolation and limitations in the depth of the soil profile; for example, Allen [1990] reports that microlysimeters are unreliable for measuring soil evaporation under crops after heavy rains.

The eddy correlation technique [see Kanemasu et al., 1979; Businger, 1986; Baldocchi et al., 1988] is a micrometeorological method that allows us to examine soil/ atmosphere exchange processes directly, in situ, and with minimal impact on the local environment. Other advantages of the eddy correlation method are its ability to measure areally averaged fluxes and to probe processes that would otherwise be missed by using chambers (such as ambient pressure fluctuations).

Preliminary eddy correlation studies of soil-atmosphere gas exchange have been conducted below forest canopies [Denmead, 1984; Baldocchi et al., 1986]. Although the eddy correlation technique has much potential for studying soillitter-atmosphere gas exchange below plant canopies, the application of this method has not been fully evaluated for use below a forest canopy. Errors can arise when the assumptions on which the technique is based are compromised or violated [see Tanner, 1967; Kanemasu et al., 1979; Businger, 1986; Baldocchi et al., 1988].

In this paper we describe micrometeorological studies of water vapor and CO_2 gas exchange from the soil/litter complex below a deciduous forest. The goals of this work are (1) to demonstrate the validity and applicability of the eddy correlation method below a deciduous forest canopy to measure forest floor gas exchange rates, (2) to examine the processes governing the emission of water vapor and CO_2 from a deciduous forest floor, and (3) to document diurnal and seasonal trends in the exchange rates of these gases. The deposition of ozone and SO_2 at the forest floor were also measured and are discussed in a companion paper (T. P. Meyers and D. D. Baldocchi, unpublished manuscript, 1991).

MATERIALS AND METHODS

We report on two eddy flux studies. The first study was conducted between April 29 and May 10, 1988. This period corresponds with the late-spring, leaf expansion phenoseason. The second study was performed between September 14 and 30, 1988. This period corresponds with the late summer to early autumn, full-leaf phenoseason. The discussion of these data is supplemented with additional measurements taken during a study performed between July 24 and August 10, 1984 [Baldocchi et al., 1986], which coincided with the summer full-leaf phenoseason.

Site Characteristics

The experiments were conducted on the U.S. Department of Energy Reservation near Oak Ridge, Tennessee (latitude 35°57'30"; longitude 84°17'15"; altitude 365 m above sea level (asl)). The field site is on a ridge in undulating terrain; a topographical map of the site is shown in Hutchison and Baldocchi [1989]. The soil is a cherty silt loam (Typic Paleudult). Its bulk density in the upper 10 cm layer is 1.33 $g \text{ cm}^{-3}$ [Luxmoore et al., 1981]. The research site and surrounding area are forested with an uneven-aged stand of oak, hickory, and loblolly pine (spp. Carya, Quercus, Pinus) representative of the Appalachian region. The mean canopy height is about 23 m. The leaf area index (LAI) and plant area index (PAI) of the forest are 4.9 and 5.5, respectively [Hutchison et al., 1986], during years of sufficient precipitation. The 1988 experimental period coincided with a severe multiyear drought; the total precipitation between the start of the winter soil recharge period (November 1, 1987) and the end of the first (June 1988) and second (October 1988) experiments were 617 mm (243 mm below average) and 1027 mm (256 mm below average), respectively. This drought caused the canopy foliage to be much sparser than in wetter years. PAI of the canopy was estimated during these experiments using an inversion of the Beer's law for radiation

penetration through a plant canopy; the solar radiation measurement system is described in *Baldocchi et al.* [1984] and the extinction coefficient, used to compute PAI, is derived from that study. PAI was computed to be 3.2 during the first study and was estimated to be 4.1 during the second study. Additional information on the vegetation, canopy, climate, and soils at the field site are presented in *Hutchison et al.* [1986], *Hutchison and Baldocchi* [1989], and Johnson and van Hook [1989].

The interpretation of the CO₂ efflux data requires information on the amount and composition of organic matter in the litter and soil. The litter biomass of a chestnut oak forest on the Walker Branch Watershed varies between about 5 Mg ha⁻¹ to about 3 Mg ha⁻¹ between April and September [*Edwards et al.*, 1989]. The O1 litter layer was about 0.01 m deep. The soil biomass at the field site is about 116 Mg ha⁻¹ [*Edwards et al.*, 1989]. The carbon/nitrogen ratio of the surface and subsurface soils on a ridge on the adjacent Walker Branch Watershed was about 20 to 1, when surveyed by Johnson et al. [1988] in 1982. The nutrient content of the O1 leaf litter is 1.24% nitrogen, 0.077% phosphorous, 0.093% potassium, and 2.11% calcium [*Edwards et al.*, 1989]. The lignin content is about 20% [Meentemeyer, 1978].

Eddy Flux Instrumentation and Data Acquisition

Three-dimensional wind velocities were measured with a sonic anemometer (model BH-478B/3; Applied Technology, Boulder Colorado). The instrument provides a 20-Hz output, with each reading equal to the average of 20 pulses. The anemometer can resolve wind velocity fluctuations of 0.0024 m s⁻¹. The path length to transducer diameter ratio (l/d) for this instrument is 25:1. Wind tunnel studies show that transducer shadowing errors on vertical velocity measurements are relatively small due to the sensor's symmetric and aerodynamic head design [Conklin et al., 1988]. The sonic anemometer boom extended 3 m ahead of the 0.75-m-wide supporting tower to minimize tower-wake effects. The anemometer boom was placed 2 m above the ground, and its azimuth orientation was 210°. We found from experience that the measured wind statistics are unreliable when the mean longitudinal wind velocity (u) is less than 0.1 m s^{-1} , $\langle w'u' \rangle > 0$, signal spikes are noted, winds are directed from beside or behind the anemometer array, or the drag coefficient (u^*/u) exceeds 0.5. Data from such periods were omitted.

A krypton hygrometer (model KH20; Campbell Scientific, Logan, Utah) was used to measure water vapor density fluctuations [see *Campbell and Tanner*, 1985]. The path length was 9 mm, and the response time was less than 0.1 s. The sensor was placed 0.2 m behind the sphere of the anemometer probe. Data were omitted when the mean voltage output was less than 0.1 V, since the analog to digital converter could not resolve all humidity fluctuations; these extreme humidity conditions only occurred during the second study.

Temperature fluctuations were measured with a microbead thermistor (Victory Engineering Corp., Springfield, New Jersey, model E45A401C). The sensor time constant is 0.072 s, and the transducer sensitivity is about 17 C V⁻¹. The sensor was placed next to the sphere of the anemometer probe.

An open-path CO_2 sensor was used to measure mean CO_2

density fluctuations during the first study, until it was damaged during a thunderstorm on day 125. The sensor was developed and built at the Lawrence Livermore National Laboratory and is described by Bingham et al. [1978]. This sensor operates as a dual wavelength, infrared absorption spectrometer. The strong CO₂ absorption band is centered at 4.27 μ m, and the reference band, one of minimal absorption, is centered at 3.85 μ m. Infrared absorption is sensed with a lead-selenium photon detector. The temperature of the detector is controlled and held at -30° C, using Peltier cooling; this feature minimizes noise, increases accuracy, and reduces drift. Two mirrors are spaced 0.25 m apart to define the sample volume. An infrared beam is reflected three times, allowing absorption to occur over a 1.5-m pathlength. The time response is less than 0.1 s. The operational sensitivity was 75 mg CO₂ m⁻³ V⁻¹, and the signal noise level was about 1.5 mg m⁻³. The sensor was placed 0.5 m behind the sphere of the anemometer probe.

A closed-path CO₂ sensor (Infrared Industries model IR 703 gas analyzer, Santa Barbara, California) was used during the remainder of the first study and during the entire second study. This sensor also operates as an infrared absorption spectrometer and implements a solid state, lead-selenium photon detector. The sensor operated on the 0- to 500-ppm scale, and its sensitivity was about 600 mg m⁻³ V⁻¹. The measurement cell is 0.175 m long, with an internal volume of 0.0886 L. The accuracy and noise are 1% of full scale. Air was sampled through a tube (Teflon, 4.5 mm ID), whose opening was placed inside the sphere of the sonic probe. The flow rates were kept moderate to prevent unwanted pressure fluctuations inside the measurement cell. In the first study the sampling tube was 3.4 m long and introduced a 3-s lag. In the second study the sampling tube was 2.5 m long and introduced a 1.75-s lag. Computer software compensated for these time lags during data acquisition and the computations of the flux covariances. The moderate flow rates introduced a moderate residence time for a fluid parcel in the measurement cell, which was 3.9 s in the fall and 4.9 s in the spring (the influence of this effect is discussed below).

The CO_2 gas analyzers were calibrated one to three times daily with standard calibration gases (Scott Specialty Gases). These gases are guaranteed to have a 1% accuracy and are traceable to the National Bureau of Standards. The calibration gases had concentrations of 314.8, 321.9, 366.2, and 381.7 ppm. Tests showed that both CO_2 instruments were linear in this range. CO_2 measurements were rejected when they periodically went off scale (exceeding 495 ppm) at night under extremely calm conditions.

The fast response instrumentation was sampled and digitized with a computer-controlled data acquisition system. The analog to digital conversion board (Data Translation, model DT2806, Marlboro, Massachusetts) has a 12-bit processor, yielding a bit resolution of 0.0048 V in the ± 10 V range.

Application of the eddy correlation technique requires that we sample with sufficient frequency and duration to avoid aliasing and to measure the spectrum of eddies that contribute to the turbulent transfer. Sensor signals were sampled at 17 Hz during the spring study and at 6 Hz during the fall study and were averaged for one-half hour periods. Evidence showing that these sampling rates were adequate to capture significant high-frequency variance and covariance contributions can be obtained by examining prior power- and



Fig. 1. The power and cospectrum of CO_2 flux measured at 2 m above the floor of a deciduous forest. Spectral densities are multiplied by natural frequency (*n*) and are normalized by the respective mean variance or covariance. These data were acquired on day 219 in 1984 between 1100 and 1130 hours.

cospectra computations. Figures 1a and 1b show that nondimensional spectral cutoff frequencies for the CO₂ power and CO₂-vertical velocity cospectra were about 2 and 1, respectively. For typical wind conditions near the forest floor ($u = 0.3 \text{ m s}^{-1}$) these nondimensional cutoff frequencies translate into dimensional values of 0.3 and 0.15 Hz, for the power and cospectra; allowing for the Shannon sampling theorem for digital conversion of analog signals, which requires that we sample at twice these cutoff values to avoid aliasing, yields cutoff frequencies that are still less than the sampling rate. A determination of proper sampling duration is more nebulous to define. On one hand, Kaimal et al. [1989] recommends that it should exceed the dominant time scale for turbulent transfer by a factor of 10. On the other hand, the sampling duration should be short enough to avoid the superposition of low-frequency diurnal trends on the time series. The dominant time scale for turbulent fluxes in the subcanopy of a deciduous forest, derived from the peak nondimensional frequency of the CO₂-vertical velocity cospectra, is of the order of 200-300 s (Figure 1b). The sampling duration used in this study was therefore adequate to measure most of the turbulent events that contribute to cospectra for the fluxes of energy, momentum, and trace gases near the forest floor.

Eddy Flux Computations

The turbulent flux density of a scalar across a horizontal plane is proportional to the mean covariance between vertical velocity (w) and scalar concentration fluctuations (c), $F = \rho_a \langle w'c' \rangle$, where ρ_a is air density, the angle brackets denotes time averaging, and primes represent fluctuations

from the mean. Turbulent fluctuations were computed as the difference between instantaneous and mean values. Mean values were computed in a real-time using a digital recursive filter [McMillen, 1988]. The filter computations employed a 200-s time constant. Tests by Verma et al. [1986] and McMillen [1988] show that little differences in eddy flux computations occur by varying the filter time constant between 100 and 1000 s.

A small nonzero vertical velocity can be introduced when the field site is not perfectly flat and when the anemometer is not perfectly aligned. We are only interested in measuring the turbulent fluxes across the mean horizontal wind streamlines, which follow the contour of the land. Therefore the three-dimensional coordinate system of the anemometer was mathematically rotated to obtain a zero mean vertical velocity and to align the longitudinal velocity (u) along the direction of the mean wind. Turbulent fluxes were then recomputed in reference to the new coordinate system [see *Wesely*, 1970].

The exchanges of sensible heat and water vapor lead to fluctuations in the density of dry air, which introduces a nonzero, mean vertical velocity that is too small to detect with the sonic anemometer [Webb et al., 1980; Leuning et al., 1982]. We applied the Webb et al. corrections to our CO_2 flux measurements. The application of these density corrections is especially critical when measuring small CO_2 fluxes, as from the soil/litter system. For example, if the corrections are omitted, we observe spurious fluxes of the wrong sign; Leuning et al. [1982] report similar findings for CO_2 exchange with and without density corrections from bare soil.

We adopt a sign convention where the turbulent flux densities, H, LE, and F_c , and the soil heat flux density (S) are positive when directed away from the surface and are negative when they are directed toward it.

Data Analysis Criteria

Micrometeorological fluxes measured at a given height are assumed to equal the exchange rates of the underlying surface as long as certain constraints relating to the conservation budget of a passive scalar are met [see Kanemasu et al., 1979; Businger, 1986; Baldocchi et al., 1988]. These constraints include (1) an extended and horizontally homogeneous, upwind surface, (2) no sources or sinks existing between the surface and the measurement level, (3) steady state conditions, and (4) level underlying surface. The following discussion confirms that the data presented in this paper generally met these constraints.

The first condition can be met if the underlying surface is uniform and level for a distance exceeding about 100 times the measurement height [Rosenberg et al., 1983; Schuepp et al., 1990]. The forest stand extends for several kilometers in all directions, so the constraint of a homogeneous underlying surface is easily met. The undulating terrain and the placement of instrument tower and laboratory facilities, however, forced us to only accept data when the mean wind came from a range of azimuth angles that were associated with flow along the dominant ridge. The terrain in the acceptable sector was relatively level, having a slope in the first 300 m upwind of less than 2°. We are confident that the undulating terrain had a minimal impact on our flux measurements. This is because (1) we rotated the wind velocity axes to allow the vertical scalar flux densities to be computed normal to the adjusted wind streamlines, (2) the wind vector rotation angles in the vertical plane were typically less than 2°, implying that the influence of the undulating terrain was small, (3) no flux divergence in tangential shear stress was measured in the stem space [Baldocchi and Meyers, 1988a], and (4) theoretical studies suggest that the impact of flow distortion by low hills on latent and sensible heat exchange are small (W. S. Weng et al., unpublished manuscript, 1991). Thereby, relaxation in the restriction regarding level terrain can be made as long as the slope is not too steep to cause flow separation and wake vortices. The forest has an observable understory canopy [Hutchison et al., 1986]. Its presence can potentially introduce errors onto our flux measurements of water vapor and CO₂ exchange if the vegetation is significantly transpiring, respiring, or photosynthesizing. The following computations, based on the conservation equation for a passive scalar, indicate that these potential sources of error were negligible. The potential error due to diffusive sources and sinks in a canopy layer (during steady state conditions) is equal to the integral, with respect to height, of the turbulent flux divergence

$$\int_0^z \frac{\partial F}{\partial z} \, dz.$$

This flux divergence can be evaluated by defining the diffusive source/sink in its resistance-analog form:

$$\frac{\partial F}{\partial z} = -a(z) \frac{\rho_x(z) - \rho_x(i)}{r_b + r_s} \tag{1}$$

where a(z) is leaf area density, $(\rho_x(z) - \rho_x(i))$ is the difference in the density of a trace gas (subscript x) measured in the air and inside the leaves, r_b is the leaf boundary layer resistance and r_s is the leaf surface resistance. Leaf area profile measurements by Hutchison et al. [1986] indicate that only 2% of the total leaf area exists below the 2-m measurement level. Using a canopy photosynthesis and stomatal conductance model [Baldocchi, 1989], we compute that net CO₂ exchange and transpiration by the understory vegetation are 0.0052 mg m⁻² s⁻¹ and 0.1 W m⁻², respectively, when incident solar radiation and atmospheric demand for water vapor were at their greatest. These theoretical exchange rates by the understory are an order of magnitude smaller than the detection limit of our eddy flux measurements (see below) and can be considered insignificant in comparison to the magnitude of the emission rates that are actually measured from the soil.

To avoid complications introduced by nonstationarity, we rejected data if the mean scalar concentration varied by more than 5% over a half-hour period. Finally, we rejected data obtained during rainy periods.

Evaluation of Flux Measurement Detection Limits and Errors

Below a forest canopy, the magnitudes of eddy fluxes are potentially small. It is thereby instructive to define the minimum resolvable flux covariances, so we can properly interpret our results. The least detectable flux is at best limited by the instrument sensitivity and the bit resolution of the analog to digital conversion board. In practice, the least detectable flux is set by the ability to detect a scalar or

 TABLE 1.
 Correction Factors for the Effects of Path Averaging, Limited Dynamic Sensor Response, and Separation Between the Measurement of Vertical Velocity and Scalar Fluctuations

<i>u</i> , m s ⁻¹	$\langle w'T' \rangle$	$\langle w'q' \rangle$	⟨w'c'⟩ Open Path	⟨w'c'⟩ Closed Path
0.1	1.037	1.138	1.271	1.115
0.2	1.040	1.141	1.275	1.179
0.3	1.039	1.139	1.276	1.233
0.4	1.036	1.137	1.273	1.278
0.5	1.033	1.133	1.266	1.321

The transfer function for the closed-path CO_2 sensor also accounts for sampling through a tube and the residence time in the measurement cell.

velocity fluctuation that exceeds the random noise of the instrument. The minimum detectable eddy covariances of sensible heat and of CO₂, using the open and closed path instruments, were about 0.45 W m⁻², 0.0072 mg m⁻² s⁻¹, and 0.029 mg m⁻² s⁻¹, respectively. The smallest resolvable latent heat flux depends on the mean voltage of the krypton hygrometer; the minimum resolvable flux ranged between 0.45 and 4.5 W m⁻² as the mean voltage output decreased from 1 to 0.1 V.

An evaluation of errors attributed to an eddy flux system is not straightforward, since the measurements are susceptible to errors from many sources. The most common errors are due to (1) a finite sampling period and sampling rate, (2) a limited dynamic instrument response, (3) nonzero separation distances between the vertical velocity and scalar fluctuation measurements, (4) a finite size of the instrument sensing path, and (5) sampling concentrations through a tube. The ramification of these undesirable effects is an underestimate of the flux by filtering or smearing turbulent eddies at particular scales [Moore, 1986; Philip, 1963; Leuning and Moncrieff, 1990]. Since we are applying our flux system under a new set of environmental conditions (low wind speeds and fluxes dominated by intermittent turbulent events), it is instructive to evaluate the magnitude of errors that may arise with this particular experimental design. Table 1 presents spectrally integrated correction factors for turbulent fluxes, according to different wind speed classes. These correction factors were calculated by weighting the appropriate spectral model of Kaimal et al. [1972] by appropriate transfer functions [see Moore, 1986; Philip, 1963]. The theoretical errors in measuring sensible heat flux density $(\rho_a C_p \langle w'T' \rangle)$, latent heat flux density, and CO₂ flux densities (with the open and closed path configurations) are less than 4%, 14%, 27%, and 32%, respectively, for the range of wind conditions typically encountered.

We chose not to apply these correction factors to our measurements, since the spectral models of *Kaimal et al.* [1972] are not representative of the turbulence regime below a forest canopy (see Figure 2). Inside a forest canopy, work by mean and turbulent kinetic energy against drag imposed by the foliage filters out many of the high-frequency turbulent fluctuations [*Baldocchi and Meyers*, 1988b]. The corrections presented in Table 1 overestimate the actual errors that are encountered in the field because the transfer function corrections are greatest at frequencies which contribute little to the turbulent fluxes, as is evident by comparing Figures 1 and 3.



Fig. 2. A comparison of the power spectrum for vertical velocity fluctuations made in the trunk space of a deciduous forest and over a flat landscape [Kaimal et al., 1972]. The shaded area defines the range of values from our field measurements. Spectral densities are multiplied by natural frequency (n) and are normalized by the w variance.

Ancillary Environmental Measurements

Ancillary measurements of solar, net, and photosynthetically active radiation, air temperature, and humidity were measured above the canopy and in the subcanopy to aid our analysis of the abiotic control of evaporation and CO_2 efflux. Short-term, point radiation measurements below a forest canopy do not represent the radiation balance that is driving the upwind energy exchanges at the forest floor because the radiation regime near a forest floor exhibits great spatial variability [*Black and Kelliher*, 1989]. Representative mea-



Fig. 3. The spectral transfer functions for the flux density measurements of sensible heat $\langle w'T' \rangle$, water vapor $\langle w'q' \rangle$, and CO₂. These data apply to flux measurements made at 2 m above a forest floor when wind speed is 0.1 m s⁻¹. The transfer functions are reported in *Moore* [1986].



Fig. 4. The sum of sensible (H) and latent heat (LE) flux densities, measured with the eddy correlation method, as a function of available energy (net radiation minus soil heat flux densities, $R_n - S$). The coefficient of determination (r^2) equals 0.752, the linear regression is 0.847, and the regression intercept is not significantly different from zero. These data consist of eddy correlation flux measurements made above the forest floor during the spring and fall experimental periods.

surements of radiation components near a forest floor can be obtained by using a moving sensor or by implementing a statistically adequate number of sensors. We were unable to make real-time net radiation measurements with our moving tram system during this experiment because its operation induces electronic interference on the circuitry of the sonic anemometers. Areally averaged radiation flux densities near the forest floor were consequently computed from Beer's law. The net radiation attenuation coefficient was determined from prior experiments with a moving tram radiation measurement system [Baldocchi et al., 1984]. During the spring study the net radiometer above the canopy malfunctioned. Missing data were computed from a regression of net radiation on solar radiation (R_g) measurements that was determined using unpublished data from a previous experiment: $R_n = 0.83 * R_g - 29$ (W m⁻²).

Litter temperature was measured with three thermistor probes. The probes were inserted 0.01 m into the litter mat. We attempted to measure soil heat flux with three sensors. These were buried below the litter mat (they were placed at 0.01 m) and were in good thermal contact with the soil. Unfortunately, two sensors were damaged by lightning, so only one instrument was available for measuring soil heat flux during the fall study. During the spring study, soil heat flux was estimated using a regression on the net radiation balance at the forest floor. The signals from these instruments were sampled and digitized on a Campbell Scientific (Logan, Utah) CR-21X data logger at 1 Hz.

Static pressure fluctuations were measured during the spring study. Atmospheric static pressure fluctuations are typically very small, being of the order of 10 Pa [Sigmon et al., 1983], so great care must be taken to ensure that they are properly measured. We used a system that was adapted from

designs developed by Elliott [1972] and Sigmon et al. [1983]. The system consisted of a differential pressure transducer (Datametrics type 590, Willmington, Massachusetts), a reference volume and a power supply. The transducer, reference volume, and plumbing were contained in an insulated box that was buried in the soil. The power supply was in an external box to avoid heating of the transducer system. The pressure transducer noise was 0.001% of full scale (0.1 Pa). Its resolution was 0.1 Pa, and its time response was 0.008 s. One end of the transducer was ported to the atmosphere. This port was 2.38 mm in diameter and was placed level to the ground surface. The Reynolds number at the port was less than 15, causing errors in static pressure measurements due to streamline curvature, standing eddies, and a Pitot effect to be negligible [see Franklin and Wallace, 1970]. A 1.58 mm ID stainless steel tube connected the port to the tap on the pressure transducer; its length was 26 cm. The reference side of the transducer was connected to a 0.7-L flask. The pressure in the reference flask was allowed to respond to unwanted low-frequency pressure variations via a small opening to the atmosphere regulated by a needle valve. The leaking reference volume acted as a high-pass filter with a time constant of 400 s. Data were sampled and digitized at 5 Hz with a Campbell Scientific 21× data logger. Variances in static pressure fluctuations were computed each half-hour. The pressure system was able to measure static pressure fluctuations ranging between 0.0025 to 2.5 Hz with minimal attenuation.

RESULTS AND DISCUSSION

Energy Balance Closure

How valid and accurate are the fluxes measured below a forest canopy with the eddy correlation method? This question can be answered by demonstrating closure of the surface energy budget. The available energy (net radiation flux density (R_n) minus the soil heat flux density (S)) must equal the sum of sensible (H) and latent (LE) heat flux densities. Figure 4 shows the relationship between H + LE and $R_n - S$. Although run-to-run variability is evident, the correlation coefficient is high (0.87) and a paired Student's *t*-test reveals that the dependent and independent variables are not significantly different from zero on the 5% probability level (Table 2). Hence we conclude that eddy flux measure-

TABLE 2. Mean Values and Standard Errors of the Energy Balance Components

Variable	Spring/Fall 1988
n	. 238
$H, W m^{-2}$	11.7 ± 1.30
LE, W m ⁻²	10.21 ± 0.80
$S, W m^{-2}$	12.4 ± 1.25
$Rn, W m^{-2}$	36.4 ± 3.16
Energy balance residual (Rn -	0.88 ± 1.01
LE - H - S, W m ⁻²	
$H + LE, W m^{-2}$	21.90 ± 1.94
$R_n - S, W m^{-2}$	22.78 ± 1.99
Student's t statistic	-0.863
t _{0.05}	1.645

n is number of half-hour periods. The Student's t statistic tests whether or not the mean energy balance residual is significantly different from zero on the 5% probability level.

ments from a deciduous forest floor are valid and reliable if data from many runs are lumped together and averaged. The scatter of data from individual half-hour periods suggests that the eddy correlation technique is accurate, but lacks precision when used near the forest floor. The scatter in Figure 4 is comparable to that observed in energy balance studies conducted above vegetation canopies [Verma et al., 1986]. This scatter is due to a combination of factors: geophysical variability and sampling and measurement errors. The spatial variability in the radiation field below a forest canopy is one source of variation; the spatial coefficient of variation of light measured near the forest floor can approach 100% over a 30-m path (D. D. Baldocchi et al., unpublished data, 1990). Natural variability in the radiative flux densities that comprise the net radiation flux density will place a constraint on the ability to close the surface energy balance in the short term. Spatial variability in the radiation field will also introduce sampling errors onto our estimate of soil heat flux density, since we do not have enough instruments to obtain a proper spatial average of this quantity. Improvements in precision of the energy balance closure can be obtained if the sampling protocol for R_n and S corresponds with the flux footprint; this footprint can be computed using algorithms discussed by Schuepp et al. [1990]. Sampling errors can also be reduced by implementing longer sampling periods, which reduce run-torun variability [see Kaimal et al., 1989]. The effectiveness of this operation is evident by the null energy balance residual that occurs when all the data are pooled (Table 2).

Processes Governing Energy Balance Exchanges

How well the evaporation from a surface is coupled to the driving potential of its external environment can be diagnosed by implementing the "omega" theory of *Jarvis and McNaughton* [1986]. Evaporation from a surface is well coupled to the product of the vapor pressure deficit and the surface conductance when omega approaches zero. Surface evaporation is decoupled from the driving potential of the atmosphere when omega approaches 1. Under this limit, *LE* is governed primarily by available energy and proceeds at values approaching its equilibrium rate.

The Omega coupling coefficient is described by

$$\Omega = \frac{s + \gamma}{s + \gamma + \gamma(g_{av}/g_s)}$$
(2)

where s is the slope of the temperature-dependent saturation vapor pressure curve, γ is the psychrometric constant, g_{av} is the boundary layer conductance to water vapor transfer

$$1/g_{av} = \left[\left(\frac{u}{u^{*2}} + \frac{2}{ku^*} \left(\frac{Sc}{Pr} \right)^{2/3} \right]^{-1}$$

u is wind speed, u^* is friction velocity, *k* is von Karman's constant (0.40), *Sc* is the Schmidt number, *Pr* is the Prandtl number, and g_s is the surface conductance.

An evaluation of omega requires that we first evaluate the surface conductance, which can be computed from the Penman-Monteith equation:

$$LE = \frac{sA + \rho C_p Dg_{av}}{s + \gamma + \gamma (g_{av}/g_s)}$$
(3)

where ρ is air density, C_{ρ} is the specific heat of air, and D is the saturation deficit $(D = q_s(T_a) - q)$, where q is the water



Fig. 5. The response of latent heat flux density (LE) measured at the floor of a deciduous forest to available energy $(A = R_n - S)$.

vapor mixing ratio and q_s is the saturation value that occurs at air temperature, T_a . Surface conductances from a dry forest floor during daytime periods typically ranged between 0.33 mm s⁻¹ and 2 mm s⁻¹, suggesting that the soil/litter surface imposes a significant but not an infinite resistance to water transfer. These soil resistances are in relative agreement with the findings of *Kelliher and Black* [1986] and *Denmead* [1984] for the forest floors of Douglas fir (*Pseudotsuga menziesii*) and *Pinus radiata* stands, respectively. We also note that these surface conductances are less than those for this forest canopy, which range between 5 and 10 mm s⁻¹ [Verma et al., 1986].

The mean daytime omega value for evaporation from the dry forest floor was 0.25 ± 0.20 . In contrast, omega values for evaporation from this deciduous forest canopy range between 0.35 and 0.65 [Verma et al., 1986]. Based on the low mean omega value, we can expect evaporation from the forest floor to be weakly dependent on available energy and relatively coupled to the atmospheric driving potential. Below, we examine this postulate in detail.

Evaporation from soil with a dry litter surface is well correlated to $(r^2 = 0.80)$, but is weakly driven by available energy ($A = R_n - S$; Figure 5). LE from the forest floor never exceeds 40 W m⁻², in spite of A being more than twice that value. On average, LE from a dry forest floor is about 43% of the equilibrium value $[LE_{eq} = (s/s + \gamma)A]$. When the litter surface is wet, LE values exceed 40 W m⁻² and are much more sensitive to changes in available energy. Under wet conditions, soil evaporation rates approach and even exceed rates of equilibrium evaporation. These results are consistent with the findings of (1) Kalma and Stanhill [1969], studying evaporation from the floor of a recently irrigated orange grove, (2) Denmead [1984], measuring evaporation from the floor of a Pinus radiata forest, (3) Walker [1984], studying evaporation under a corn canopy, and (4) Kelliher and Black [1986], studying evaporation from the floor of a Douglas fir stand.

It may seem surprising to those who have spent much time walking through a forest that evaporation from its floor is



Fig. 6. Schematic showing the analytical solution for the time rate of change of latent heat flux above a forest floor.

more closely coupled to the saturation deficit of the external environment than is evaporation from the overlying and aerodynamically rougher forest canopy. Yet, an accumulating body of data suggests that evaporative fluxes from many forest floors and understory vegetation are, indeed, well coupled to the atmospheric deficit of the ambient air [Finnigan and Raupach, 1987; Black and Kelliher, 1989; Kelliher et al., 1990]. This close coupling is made possible by large-scale eddies that have sufficient energy to penetrate deep into the canopy and, thereby, dominate the transfer of trace gases at this locale [Denmead, 1984; Denmead and Bradley, 1985; Baldocchi and Meyers, 1988a, b]. These large-scale eddies are identified by a quasi-periodic sequence of events that occur every 50-100 s [Gao et al., 1989; Bergstrom and Hogstrom, 1989]. First, air rapidly sweeps into the canopy trunk space. This sweep is then followed by an ejection of air out of the canopy and a relative quiescent period.

Equilibrium LE is not attained for understory evaporation when the litter surface is dry because intermittent turbulence displaces the air in contact with the evaporating surface before it attains equilibrium. An analytical expression can be derived to demonstrate this point quantitatively. Consider the exchanges of latent and sensible heat into a conceptual "box" located in the canopy stem space. This box is assumed to be decoupled from the overlying atmosphere during quiescent periods by the overlying canopy crown, so sensible and latent energy fluxes at the top of the box are assumed to be negligible. Furthermore, latent and sensible energy fluxes into the sides are assumed to equal those leaving. LE at the soil surface is estimated with the Penman-Monteith equation (4), where only the saturation deficit (D)depends on time. The time derivative of LE during "calm" periods can thereby be defined as

$$\frac{dLE(t)}{dt} = \frac{dLE(t)}{dD(t)} \frac{dD(t)}{dt} = \frac{g_{av}[sA - (s+\gamma)LE(t)]}{h[s+\gamma+\gamma(g_{av}/g_s)]}$$
(4)

The time rate of change in D is due to the fluxes of latent and sensible heat from the soil/litter surface. It equals

$$\frac{dD(t)}{dt} = \frac{sA - (s + \gamma)LE(t)}{\rho C_p h}$$
(5)

where h is the height of the box (we assume a value equal to the measurement height, 2 m). Solving the differential equation for LE yields

$$LE(t) = \frac{s}{s+\gamma} A + \exp\left(-\frac{t}{\tau}\right) \left[LE(0) - \frac{s}{s+\gamma} A \right]$$
(6)

where τ is the time constant:

$$\tau = \frac{h[s + \gamma + \gamma(g_{av}/g_s)]}{g_{av}(s + \gamma)}$$
(7)

Equilibrium evaporation $(s/s + \gamma) A$ is approached when $t \gg \tau$.

Figure 6 shows how *LE* in the canopy trunk space varies with time for a given set of initial conditions. Equilibrium evaporation rates are approached as time exceeds thrice the time constant. The actual value of the evaporation time constant (9) depends on g_{av} and g_s (Figure 7). When the soil/litter surface is wet, g_s approaches infinity, causing the



Fig. 7. Analytical calculations of time constants for latent heat transience as a function of forest floor boundary layer conductance.



Fig. 8. The dependence of the forest floor evaporation partitioning of the ratio (LE/R_n) on static pressure fluctuations.

time constant to range between 100 and 200 s for the typical values of g_{av} that are encountered in the field. Time constants of this magnitude enable soil evaporation to theoretically approach equilibrium levels during the period between coherent turbulent events, which is consistent with the results shown in Figure 5. On the other hand, when the soil/litter surface is dry, the time constant ranges between 1500 and 5000 s, for typical values of the soil/litter surface conductance. The renewal of ambient air in the canopy trunk space by intermittent turbulence is too rapid for the understory to permit equilibrium rates of evaporation in this case.

Our finding that mean soil/litter evaporation from a dry surface operates well below the rate of equilibrium evaporation is consistent with computations by Finnigan and Raupach [1987], who show that intermittent turbulence can prevent leaves from evaporating in a closed plant canopy understory at equilibrium levels. Kelliher and coworkers [Kelliher and Black, 1986; Kelliher et al., 1990], on the other hand, report that evaporation from understories of open forest stands exceed equilibrium rates by factors of 1.43 (Pinus radiata) and 2.0-2.7 (Pseudotsuga and menziesii). Kelliher et al.'s data represent a converse situation. Their measurements were made under sparse overstory canopies and the understories were vegetated, so their canopies had much lower surface resistances and a greater evaporating surface area than the soil/litter surface under a denser deciduous forest canopy.

Static pressure fluctuations have virtually no role in governing the partitioning of net radiation into forest floor evaporation from dry surfaces (Figure 8). The pathway through which vapor travels from its source to the atmosphere is quite tortuous when the soil is dry and probably prevents pressure fluctuations from impacting on its transfer. However, when the litter surface is wet, the partitioning of R_n into *LE* is strongly correlated with static pressure fluctuations ($r^2 = 0.66$). Pressure fluctuations probably enhance soil-atmosphere gas exchange by "pumping" water vapor out of litter and soil pores. This result is supported by the laboratory studies of *Kimball and Lemon* [1971]. They show that static pressure fluctuations increase the efflux of heptane from coarse-grained soils when the source is located



Fig. 9. (a) The diurnal variation in net radiation, latent heat, sensible heat, and soil heat flux densities near the forest floor. (b) The diurnal variation in litter temperature, air temperature, and vapor pressure deficit above a deciduous forest floor. These data are averaged from the measurements made during the spring study.

near the surface, which is analogous to evaporation from loose, wet litter. We cannot discount the role of a reduced boundary layer resistance being coupled with increasing σ_p , since the pressure and wind fields are closely linked [see Wyngaard, 1982; Shaw et al., 1990]. However, the differential dependence of *LE* on σ_p over wet and dry litter regimes definitely supports the hypothesis that σ_p can independently modulate soil/litteratmosphere exchange during wet conditions.

Diurnal Variations: Energy Exchanges

Figure 9a shows the mean diurnal pattern of R_n , LE, H, and S during the spring study; the mean diurnal courses for soil and air temperature and vapor pressure deficit are shown in Figure 9b. All fluxes are negligible at night. Near sunrise, R_n increases rapidly and attains a maximum near noon. Since the canopy is not closed, appreciable levels of R_n are observed near the forest floor throughout the daylight period. Midday values exceed 125 W m^{-2} , which are twice the values typically reported near the floor of closed forest canopies [Baldocchi et al., 1986; Black and Kelliher, 1989]. The diurnal patterns of LE and S are in phase with that of R_n , while the daily course of H lags R_n in the morning and follows it in the afternoon. Mean midday values of H and Sexceed 40 W m⁻², while the maximal value for mean LE is capped at about 25 W m⁻². The midday *LE* values resemble those measured during our 1984 summer study [Baldocchi et al., 1986], while, in contrast, H values were near zero during



Fig. 10. The diurnal variation in net radiation, latent heat, sensible heat, and soil heat flux densities near the forest floor. These data are averaged from measurements made during the autumn study.

the 1984 summer study. Comparing midday H values with the differences in air and litter/soil temperatures shows an occurrence of countergradient transport; sensible heat flux is directed upward, while the mean temperature gradient suggests otherwise. Countergradient transport is a real phenomenon that has been documented inside plant canopies experimentally [Denmead and Bradley, 1985; Baldocchi and Meyers, 1988a, b] and theoretically [Finnigan and Raupach, 1987]. It is an artifact of nonlocal, large-scale turbulent events contributing disproportionately to mass transfer while near-field diffusion contributes disproportionately to the concentration field.

The magnitude of the mean midday R_n near the forest floor was greatly reduced during September (Figure 10). This result occurred because the canopy LAI was greater and intercepted more incoming solar radiation. In spite of reduced available energy maximum, *LE* values of about 25 W m⁻² are still achieved. Compensation for reduced R_n occurs through reductions in the magnitudes of *H* and *S*. The constancy of mean maximum *LE* reaching about 25 W m⁻² near the forest floor over a wide range of conditions has been explained above.

CO_2 Efflux

Instrument intercomparison. The two CO_2 sensors used in this study have different design features and operate differently, so it is instructive to first compare their performance characteristics. Figure 11 shows normalized power spectra for the open- and closed-path CO_2 sensors; the data are from a prior unpublished study made above the canopy in 1984. The open-path CO_2 sensor exhibits near-ideal performance characteristics. A prominent peak is observed at nondimensional frequencies (f) ranging between 0.001 and 0.01. At frequencies exceeding the peak, the normalized power spectrum densities decrease with a slope on f ap-



Fig. 11. A comparison of the power spectra of CO_2 fluctuations measured with an open-path and closed-path CO_2 analyzer.

proaching -2/3, which is characteristic of inertial subrange turbulence [Stull, 1988]. Only at very high nondimensional frequencies (f > 2.5) does sensor performance deviate from ideal; at these frequencies the finite sampling volume of the sensor begins to distort the spectrum. The spectral characteristics of the open-path sensor resemble those reported by Anderson and Verma [1985] and Anderson et al. [1986] for an identical sensor. In contrast, the power spectrum of the closed-path sensor has an inertial subrange slope of 1. This feature indicates that its signal is contaminated by white noise [Wesely and Hart, 1985]. White noise will not significantly affect the accuracy of flux measurements as long as the noise is not correlated with vertical velocity fluctuations, which can be assumed in the inertial subrange where turbulence is locally isotropic. High-frequency white noise errors can be reduced by subjecting the instrument to lower frequency turbulence, as by raising the measurement height. The greatest limitation in using the closed-path CO₂ sensor, or any noisy instrument, is its lower precision. This artifact primarily increases run-to-run variability, which can be overcome by lumping and averaging data from many periods [Weselv and Hart, 1985].

We were unable to compare the performance of the openand closed-path CO₂ sensors side-by-side. Instead, we compared the two sensors by examining their performance under similar environmental conditions. Figure 12 shows that the two sensors yield CO₂ flux densities that respond to temperature in a similar manner. The closed-path sensor produces a biased estimate of the CO₂ efflux, compared to that from the open-path sensor, but it is small (<0.02 mg m⁻² s⁻¹).

Processes Governing CO₂ Efflux

Edwards et al. [1989] state that temperature is the primary regulator of forest floor respiration over a broad range of soil moisture conditions. Temperature is expected to strongly control CO_2 efflux rates because it governs the rates of the chemical reactions that are associated with root respiration and microbial activity. However, we find that litter temperature accounts for only 44% of the variance in CO_2 efflux rates measured with the open-path sensor during the spring experimental period (Figure 12). Litter temperature is a fair



Fig. 12. The response in CO_2 efflux from a deciduous forest floor to litter temperature. These data are from the spring period.

indicator of CO_2 efflux, since the respiratory activity of roots and microbial populations in deeper layers is comparatively reduced due to somewhat colder temperatures in those zones. It also is important to recognize that the spatial variability in CO_2 emission from forest soils can be great. *Cropper et al.* [1985] report spatial coefficients of variation exceeding 60%, so the scatter in Figure 12 is in part reflecting this natural variability. The coefficient of determination (r^2) would have probably been greater if we had a more representative spatial estimate of litter temperature (both horizontally and vertically). The eddy correlation method measures CO_2 fluxes from an upwind area of many hundred square meters [*Schuepp et al.*, 1990], while the estimate of litter temperature was derived from three probes randomly distributed in a 10 m² horizontal area.

Cool litter temperatures (below 14°C) restrict efflux values below 0.05 mg m⁻² s⁻¹, whereas warmer temperatures yield a nonlinear increase in efflux rates, with maximum values approaching 0.25 mg m⁻² s⁻¹. The magnitude of the CO₂ efflux rates and the shape of the temperature response curve (and its associated scatter) closely resemble data reported by *Reiners* [1968] from a Minnesota oak forest and measured with chambers. Our CO₂ efflux measurements are also within the range of values measured from other nearby deciduous forest floors with chamber methods [*Edwards et al.*, 1989; *Larkin and Kelly*, 1987]. On the other hand, these values are considerably less than values we measured in a prior study during the summer period [*Baldocchi et al.*, 1986], which experienced warmer soil temperatures and no drought.

 CO_2 efflux rates increase markedly with increasing levels in the standard deviation in static pressure (σ_p) (Figure 13); $r^2 = 0.56$. Static pressure fluctuations aid the diffusion of gas through coarse soils and loose litter through pumping action [Kimball, 1983; Kimball and Lemon, 1971]. Since these data are the first, to our knowledge, to document a role for pressure fluctuations on regulating forest floor CO_2 exchange, we must interpret them with caution. Litter temperature and σ_p are also well correlated (r = 0.70); σ_p increases in response to development of the convective, planetary boundary layer, which is due to sensible heat flux from the warming surface [Stull, 1988]. However, we argue that the



Fig. 13. The response in CO_2 efflux from a deciduous forest floor to static pressure fluctuations. These data are from the spring study.

observed role of σ_p on F_c is probably not an autocorrelative effect of litter temperature because σ_p explains a greater amount of the variance in F_c than does litter temperature and because there is a causal mechanism for σ_p to act on F_c . Furthermore, since we independently demonstrated that pressure fluctuations enhance the efflux of water vapor from wet litter, it is plausible to conclude that the data in Figure 13 suggest that σ_p is also modulating CO₂ exchange from the litter.

A definite diurnal course in CO_2 efflux is observed in the spring (Figure 14*a*). This diurnal course is in phase with the daily patterns of litter temperature (Figure 14*b*) and vertical velocity fluctuations (Figure 14*c*) (σ_w is a surrogate for σ_p ; Figure 15). All three variables are small at night and increase at daybreak until a midday maximum. Efflux values are low at night due to cooler temperatures and reduced turbulence (which results from the stable thermal stratification of the atmospheric surface layer). Turbulence and temperature increase during the day due to surface heating. The buildup of the convective, planetary boundary layer generates turbulence, and surface heating increases respiratory activity: two modes of action that promote the transfer of CO_2 more effectively between the soil surface and the atmosphere.

The diurnal course shown here differs somewhat from the diurnal course that we published for the 1984 study [Baldocchi et al., 1986]. In our previous study we noted an increase in soil/litter respiratory activity near sunset [Baldocchi et al., 1986, Figure 2], and attributed it to rapid catabolism of translocated carbohydrate. In retrospect, we offer a more plausible explanation for those observations. Near sunset the thermal stability of the overlying atmosphere often changes rapidly from an unstable to a near-neutral or stable regime (the sign of H changes from positive to negative; Figure 9). This reduces the turbulent mixing inside the canopy and causes CO₂ concentrations in the trunk space to build up quickly. Under such nonsteady conditions, our CO₂ flux measurements are not constant with height and must be discarded, a criterion that we should have exercised more strictly in the 1984 study. In the 1984 study [Baldocchi et al., 1986, Figure 3], we also reported that CO_2 effluxes were



Fig. 14. The diurnal course in CO_2 efflux, litter temperature, and fluctuations in vertical velocity during the spring study.



Fig. 15. The response of vertical velocity fluctuations to static pressure fluctuations.



Fig. 16. The response in CO_2 efflux from a deciduous forest floor to litter temperature (a) and (b) the response to vertical velocity fluctuations. These data are from the autumn period.

diminished in the late afternoon, but had no firm explanation for our observation. Based on the data in Figure 14, it seems that late afternoon reductions in turbulent fluctuations act to inhibit CO_2 efflux, even though soil temperatures remain relatively warm during this period.

Figure 16 shows that CO_2 efflux rates during the autumnal study are much smaller than values measured during the spring, even though autumnal soil and litter temperatures were warmer and the soil and litter were relatively wet during the experimental period. Autumnal CO_2 efflux rates are also poorly correlated to litter temperature (Figure 16*a*) and to fluctuations in vertical velocity (Figure 16*b*) (a surrogate for pressure fluctuations).

The magnitude of the autumnal CO_2 efflux rates is much smaller than values measured in nearby forests during more favorable conditions [Larkin and Kelly, 1987; Edwards et al., 1989]. The period during which our experiment was conducted was preceded by a drought and abnormally warm summer. These events probably conspired to restrict CO_2 efflux rates. First, the drought probably reduced the fine root mass in the upper soil horizons [see Kaufmann, 1990; Santantonio, 1990]; this contention is supported indirectly by the 20% reduction in LAI. The fine root mass normally contributes to over 70% of the CO_2 evolved from the forest floor in Walker Branch Watershed [Edwards et al., 1989]. Since respiration rates are a function of biomass weight [Hunt and Loomis, 1979], any reduction in root biomass will

TABLE 3. Moisture Status						
Study 1, April–May 1988						
Day Numb	er	ppt, mm	Soil Moisture, cm ³ cm ⁻³ (Neutron Probe, 15–30 cm)			
120 124 125 130		T 2.8 16.5 11.2	0.40 (before rain)			
Study 2, September-October, 1988						
Day	Precipitation, mm	Litter Moisture ((wet - dry)/wet, g g ⁻¹ , upper 1 cm)	Soil Moisture Gravimetric ((g g ⁻¹) upper 15 cm)			
260 261 263 264 265 266 267 268 269 270 271	11.1 33.0 0 0.8 0 0 7.4 0.8 0 0	0.587 0.537 0.411 0.487 0.379 0.599 0.435 0.419	0.228 0.177 0.202 0.195			
272 273	0 0.025		0.173 0.143			

reduce forest floor CO_2 efflux rates. We cannot argue that the drought limited microbial respiration in the same manner that it may have limited root respiration. Soil and litter moisture were adequate (Table 3) during the autumnal study, and manipulative studies show that the respiratory activity of microbial populations respond within hours to the reintroduction of moisture [Cook et al., 1985]. The second event restricting CO₂ effluxes involves the fact that soluble substances and nonlignified carbohydrates decompose most rapidly (and a warm summer accelerates this process), leaving litter that predominantly consists of lignified carbohydrates and decomposes slowly [Berg, 1986; Meentemeyer, 1978]. Third, reductions in litter decomposition and fine root respiration will move the location of the dominant source of CO₂ to deeper soil layers. Litter temperature would therefore not be expected to govern CO₂ efflux as well as temperature measured in deeper layers, since deeper soil layers will experience a much smaller daily variation and its daily course will be out of phase with litter temperature. Moving the dominant source layer to deeper soil levels will also decouple CO₂ efflux rates from the influence of static pressure fluctuations [Kimball, 1983; Kimball and Lemon, 1971], as is observed in the decoupling between F_c and the σ_p surrogate, σ_w .

When data from several days are lumped together and averaged, we observe a distinct diurnal course in CO₂ efflux during the fall (Figure 17a). Maximum values peaked at about 0.06 mg m⁻² s⁻¹, just after 1000 hours. The timing of this peak leads those for litter temperature (Figure 17b) and the standard deviation in vertical velocity fluctuations (Figure 17c). It is unclear from our measurements why this happened. Nocturnal efflux rates are below the detection limit of the eddy flux measurement system; they are less than

Fig. 17. The diurnal course in CO₂ efflux, litter temperature, and fluctuations in vertical velocity during the autumn study.

 $0.02 \text{ mg m}^{-2} \text{ s}^{-1}$. Reduced static pressure fluctuations and turbulence limit CO₂ efflux at night, in spite of the maintenance of moderate litter temperatures.

CONCLUSIONS

Our results suggest that the eddy correlation method can be used to make reliable flux measurements above the floor of a deciduous forest since we are able to demonstrate closure of the surface energy balance. But we emphasize that measurements obtained from a forest floor must be subjected to strict scrutiny and can be accepted only if they meet the requirements on which the technique is based; these include steady state conditions, no sources and sinks between the soil surface and measurement height, and an extended level and horizontally homogeneous upwind fetch.

The magnitude of water vapor and CO₂ fluxes near a forest floor are small. Yet the unique conditions of the forest floor microclimate work to our advantage in measuring small fluxes reasonably well. Fluxes near a forest floor are driven by low-frequency, turbulent events that have relatively large signals. Therefore they are easier to detect than turbulent events that drive small fluxes in the surface layer above a plant canopy, the latter being associated with small concentration and velocity fluctuations.

The eddy correlation method is not without flaws. One limitation of the eddy correlation technique is that it cannot make uninterrupted measurements as well as chamber methods. Data must be rejected when conditions violate the premises on which the technique is based. Consequently, we feel it is wisest to use carefully applied chamber and mi-



crometeorological methods in concert to study the processes governing fluxes near a forest floor. This combined approach would address the effects of temporal and spatial variability best.

The dry soil/litter surface exerts a significant resistance to water vapor transfer. The magnitude of the surface resistance and low wind speeds cause the time constant for a transience in latent heat exchange to be very long (exceeding 1000 s). Since the intermittent turbulent events occur on time scales of 100 s, equilibrium evaporation rates are not achieved (except when the litter surface is wet). Evaporation from dry forest floors is relatively well coupled to the driving potential of the atmosphere. The interrelation between turbulence time scales and transient time scales has great relevance in modeling the emission and deposition of chemically reactive species at the soil, such as NO, CH_4 , and O_3 . Fast reacting chemical species may never leave the trunk space volume if they are consumed in the period between intermittent gusts [Jacob and Bakwin, 1991].

 CO_2 efflux rates are governed by litter and soil temperature and by static pressure fluctuations. Yet the degree of this control changes over the year as the soil level of the dominant CO_2 source changes. Future studies of trace gas emission should implement profile measurements of soil and litter temperature and CO_2 concentration to obtain a more concrete handle on the thermal control of local sources. The strong dependence of efflux rates on pressure fluctuations suggests that pressure fluctuations may also play a significant role on modulating the emissions of other trace gases from forest soils, such as N₂O, NO, H₂S, and COS. Consequently, budget estimates derived from chamber measurements may need to be reevaluated.

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