FLUX FOOTPRINTS WITHIN AND OVER FOREST CANOPIES

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Abstract. The characteristics of turbulence within a forest are spatially heterogeneous and distinct from those associated with the surface boundary layer. Consequently, the size and probability distribution of 'flux footprints' emanating from sources below a forest canopy have the potential to differ from those observed above forests.

A Lagrangian random walk model was used to investigate this problem since no analytical solution of the diffusion equation exists. Model calculations suggest that spatial characteristics of 'flux footprint' probability distributions under forest canopies are much contracted, compared to those evaluated in the surface boundary layer. The key factors affecting the statistical spread of the 'flux footprint', and the position of the peak of its probability distribution, are horizontal wind velocity and the standard deviations of vertical and horizontal velocity fluctuations. Consequently, canopies, which attenuate mean horizontal wind speed, or atmospheric conditions, which enhance vertical velocity fluctuations, will contract flux footprint distributions mostly near the floor of a forest. It was also found that the probability distributions of the 'flux footprint' are narrower when horizontal wind velocity fluctuations are considered, instead of the simpler case that considers only vertical velocity fluctuations and mean horizontal wind velocity.

Key words: Lagrangian model, Micrometeorology, Biosphere-atmosphere interactions, Diffusion modelling, Flux footprint, Canopy turbulence

1. Introduction

The transfer of mass and energy between the biosphere and atmosphere is composed of fluxes to and from the vegetation and underlying soil. To understand fully the biological and physical processes controlling these fluxes, it is incumbent that experimentalists probe the relative contributions of vegetation and the soil systems independently. Typically, trace gas fluxes over the soil system, under plant canopies, are measured directly with chambers (Hutchinson and Livingston, 1993), mini-lysimeters (Black and Kelliher, 1989) or the eddy covariance method (Baldocchi and Meyers, 1991; Baldocchi and Vogel, 1996). Trace gas fluxes associated with the vegetative component, on the other hand, are often evaluated indirectly. One approach measures vegetative fluxes as the difference between fluxes occurring at the canopy-atmosphere interface and over the soil surface. Examples of the application of this difference-method include studies of canopy-scale transpiration (Black and Kelliher, 1989; Hollinger et al., 1994; Saugier et al., 1997), photosynthesis (Biscoe et al., 1975; Baldocchi et al., 1987; Black et al., 1996; Rochette et al., 1996; Baldocchi and Vogel, 1996), respiration (Lavigne et al., 1997), and NO_x emission and deposition (Jacob and Bakwin, 1991).

The tall stature of forests allows researchers the latitude to employ eddy covariance measurement systems over and under a forest stand. Under many circum-

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stances the application of the eddy covariance method under forests is preferred over the use of chambers. Reasons for this preference include: 1) eddy covariance measurements are continuous; 2) they are less labour intensive; 3) they have a minimal impact on the system being studied; and 4) their measurements represent a larger sampling area than is sensed by conventional chambers (see Baldocchi and Meyers, 1991; Freijer and Bouten, 1991; Hutchinson and Livingston, 1993). This last assumption has been accepted on faith and has not been subjected to theoretical or experimental scrutiny.

The size and representativeness of the upwind fetch that is sampled by an eddy covariance measurement system under a forest affects its deployment and the interpretation of the data it acquires. For example, we have attempted to validate eddy covariance measurements under forests by examining the ability to close the surface energy balance (e.g. Baldocchi and Meyers, 1991; Baldocchi and Vogel, 1996). This exercise typically involves the deployment of a track, on which radiation sensors traverse, and an array of soil heat flux plates. If the spatial field under a forest is non-isotropic and the eddy covariance system samples a smaller area than expected, several relevant questions arise. First, where should one place the radiation track and soil heat flux plates, to make a comparison with other energy balance components? And second, how long should the track be to obtain a mean representative of the footprint sensed by the eddy covariance system?

One way to examine the spatial dimension of a 'flux footprint' measured near the floor of a forest is to evaluate, numerically, the spatial probability distribution of surface sources that would be sensed theoretically at a reference tower. Over the last decade numerous investigators have applied Lagrangian (Leclerc and Thurtell, 1990; Horst and Weil, 1992; Flesch, 1996) and Eulerian (Gash, 1986; Schuepp et al., 1990; Horst and Weil, 1992; Schmid, 1994; Leclerc et al., 1997) diffusion theory to assess 'flux footprints' in the surface boundary layer *above* vegetation and across horizontal inhomogeneities (Luhar and Rao, 1994). At present, both modelling approaches are mature and have been verified for boundary-layer flow above vegetation (Finn et al., 1996; Leclerc et al., 1997). On the other hand, 'flux footprint' modelling exercises have not been applied within vegetation.

At this juncture, the results from 'flux footprint' studies conducted in the surface boundary layer cannot be applied to the circumstance below vegetation. This restriction is imposed because the velocity, time and length scales of turbulence are heterogeneous below the canopy-atmosphere interface and differ from those measured in the surface boundary layer (Raupach, 1988; Wilson, 1989; Raupach et al., 1996; Finnigan and Brunet, 1995). For example, the standard deviation of vertical velocity (σ_w) above vegetation is about 1.25 times friction velocity under neutral thermal stratification, while below canopy height, this metric decreases quasi-linearly with depth. Typical values for σ_w, near the soil and under vegetation, range between 15 and 40% of the quantity measured in the surface boundary layer (Amiro, 1990; Finnigan and Brunet, 1995; Raupach et al., 1996). The mean horizontal wind velocity (U) also decreases markedly with depth into a plant canopy

(Raupach, 1988; Raupach et al., 1996). The vertical profile of wind velocity is often modelled as an exponential function of depth into the canopy (Cionco, 1965). In contrast, the mean horizontal wind profile above a plant stand adheres to a logarithmic wind law (Kaimal and Finnigan, 1994), and is a function of the zero plane displacement, roughness length, and Obukhov length scale. The magnitude of turbulence intensity (the ratio between the standard deviation and the mean value of wind speed) typically exceeds 100% inside vegetation, which contrasts with values below 50% in the surface boundary layer (Baldocchi and Meyers, 1988; Raupach, 1988; Wilson, 1989). Finally, the integral turbulence time scale (T_L, the time over which the velocity of fluid elements remains correlated with themselves due to the persistence of turbulent motion) is generally constant within vegetation, since its constituents, the Lagrangian length scale (L_w) and σ_w, diminish with depth into the canopy (Raupach, 1988). In contrast, T_L increases with height in the surface boundary layer.

Lagrangian diffusion theory is particularly attractive for solving atmospheric flow problems that are characterized by heterogeneous turbulence. In particular, Lagrangian diffusion theory is appropriate for studying ‘flux footprint’ probability statistics because it can evaluate the movement of an ensemble of fluid elements, whose motion depends on the mean horizontal wind velocity, the standard deviations of vertical and horizontal wind velocity, and the Lagrangian time scale (Durbin, 1980; Sawford, 1985; Thomson, 1987; Flesch and Wilson, 1992; Wilson and Sawford, 1996).

The main goal of this paper is to use a Lagrangian random walk model to evaluate the effect of canopy architecture and turbulence characteristics on the ‘flux footprint’ probability density functions within and above a forest canopy. With the practical goal of designing better flux experiments in mind, the Lagrangian ‘flux footprint’ model is applied to the circumstances of two recent experiments we have conducted. One case involves a tall and dense temperate broad-leaved forest, the other involves a moderate and sparse boreal pine forest.

2. Theory

The fundamental questions being asked and addressed here are: (i) what is the upwind area sensed by an instrumented tower that is placed within a forest? and (ii) what proportion of the material or energy derives from specific upwind patches?

The definition of a ‘flux footprint’ source area and its spatial distribution requires a statistical measure of the probability that a fluid element released at point (x_r, y_r, z_r) will be measured at a position downwind from the ground source, denoted x, y, z. A mathematical description of the flux footprint source area is derived from the relationship between a flux density measured at a particular point in space, denoted $F(x, y, z_m)$, and the source probability density (or weighting) function ($f(x - x_r, y - y_r, z_m)$) of material emitted by the surface (see Schmid, 1994; Horst and Weil, 1992), viz.

$$F(x, y, z_m) = \int_{-\infty}^{\infty} \int_{-\infty}^{x} Q(x_r, y_r, 0) \cdot f(x - x_r, y - y_r, z_m)\, \mathrm{d}x_r\, \mathrm{d}y_r, \tag{1}$$

where $Q(x_r, y_r, 0)$ is the unit-area, source strength at the ground ($z = 0$). The source probability density function, $f(x - x_r, y - y_r, z_m)$, can be derived analytically or numerically, provided one has sufficient information on how fluid elements diffuse and are advected through the atmosphere. Furthermore, once the probability density function is known, the source area can be defined by integrating $f(x - x_r, y - y_r, z_m)$ with respect to x_r and y_r.

For this application, turbulent diffusion and advection was computed in the vertical and horizontal directions, above and within a plant canopy, using a Lagrangian framework (see Durbin, 1980; Thomson, 1987; Raupach, 1988; Wilson and Sawford, 1996). The vertical displacement of fluid parcels was computed as a function of time,

$$\mathrm{d}z = w \cdot \mathrm{d}t, \tag{2}$$

where w is the Lagrangian vertical velocity and dt is differential time increment. Incremental changes in vertical velocity were computed using the Langevin equation, an algorithm that is weighted by a deterministic forcing (which is a function of the fluid parcel's previous velocity) and a random forcing term (Thomson, 1987),

$$\mathrm{d}w = a(z, t, w)\, \mathrm{d}t + b(z, t, w) \mathrm{d}\xi. \tag{3}$$

The coefficients $a(z, t, w)$ and $b(z, t, w)$ are non-linear functions of w and are defined to account for inhomogeneous turbulence. The term dξ defines a Gaussian random forcing with a mean of zero and variance of dt.

The terms $a(z, t, w)$ and $b(z, t, w)$ are derived from the budget equation for the Eulerian probability density function of w (the Fokker–Planck equation) (Thomson, 1987). For the one-dimensional case, where velocity fluctuations and gradients occur only in the vertical direction, Equation (3) becomes,

$$\mathrm{d}w = \left(-\frac{w_L}{T_L} + \frac{1}{2} \left[1 + \frac{w_L^2}{\sigma_w^2} \right] \frac{\partial \sigma_w^2}{\partial z} \right) \mathrm{d}t + \sqrt{\frac{2\sigma_w^2}{T_L}}\, \mathrm{d}\xi. \tag{4}$$

This algorithm has gained acceptance, theoretically, for it meets the model criteria proposed by Sawford (1985) and Thomson (1987), including the so-called well-mixed criterion.

For the numerical calculations performed here, the finite difference algorithm of Luhar and Britter (1989) is used, which expresses the random operator in terms of an increment (i) that has a mean of zero and a variance of one (dΩ),

$$w_{i+1} = w_i + \left(-\frac{w_i}{T_L} + \frac{1}{2} \left[1 + \frac{w_i^2}{\sigma_w^2} \right] \frac{\partial \sigma_w^2}{\partial z} \right) \Delta t + \sqrt{\frac{2\sigma_w^2}{T_L} \Delta t}\, \mathrm{d}\Omega. \tag{5}$$

Equations (4) and (5) do not consider the impact of non-Gaussian and skewed statistical properties on the movement of fluid parcels, although these are distinct characteristics of canopy turbulence (Raupach, 1988; Wilson, 1989; Raupach et al., 1996). The rationale for neglecting the impact of non-Gaussian turbulence statistics on Equations (4) and (5) is based on information in Wilson and Sawford (1996). They report that Lagrangian models, which considered the effect of non-Gaussion turbulence on fluid element movement, perform worse than Thomson's 1987 model, which assumes Gaussian and Eulerian turbulent statistics.

Horizontal displacements were computed for two situations. The simplest case assumed that horizontal displacements were a function of the mean, horizontal wind velocity (U),

$$\mathrm{d}x = U(z) \cdot \mathrm{d}t. \tag{6}$$

The second, and more complex case, considered the impact of horizontal wind velocity fluctuations (u),

$$\mathrm{d}x = (U + u)\,\mathrm{d}t. \tag{7}$$

The inclusion of u in Equation (7) necessitates the evaluation of the Langevin model for horizontal wind velocity,

$$\mathrm{d}u = a(z,t,u)\,\mathrm{d}t + b(z,t,u)\,\mathrm{d}\xi. \tag{8}$$

Algorithms presented in Flesch and Wilson (1992), which were derived from the theory of Thomson (1987), were used to define the coefficients $a(z,t,u)$ and $b(z,t,u)$. When applying the Langevin equation to the case of two-dimensional turbulent transfer, the definition of the b coefficient in Equations (8) and (3) remains the same,

$$b = \sqrt{\frac{2\sigma_w^2}{T_L}}. \tag{9}$$

The $a(z,t,u)$ coefficient for situations where w and u fluctuations are considered, however, differs markedly from the corresponding term in Equation (4). Its derivation from the Fokker–Planck equation (Flesch and Wilson, 1992) yields a relationship that includes terms relating to the covariance between w and u (wu), the mean horizontal wind velocity gradient ($\mathrm{d}U/\mathrm{d}z$) and the standard deviation in horizontal wind velocity (σ_u),

$$\begin{aligned} a_u \;=\; & \frac{1}{2(\sigma_u^2\sigma_w^2 - \overline{uw}^2)} b_u^2[\sigma_w^2 u - \overline{uw}w] + \frac{1}{2}\frac{\partial\overline{uw}}{\partial z} + w\frac{\partial U}{\partial z} + \frac{1}{2(\sigma_u^2\sigma_w^2 - \overline{uw}^2)} \\ & \times \left[\sigma_w^2\frac{\partial\sigma_u^2}{\partial z}wu - \overline{wu}\frac{\partial\sigma_u^2}{\partial z}w^2 - \overline{wu}\frac{\partial\overline{wu}}{\partial z}wu - \sigma_u^2\frac{\partial\overline{wu}}{\partial z}w^2\right]. \end{aligned} \tag{10}$$

For the case of two-dimensional turbulent transfer, one must also re-evaluate the Fokker–Planck equation and define another algorithm for $a(z, t, w)$ (see Flesch and Wilson, 1992),

$$a_w = \frac{1}{2(\sigma_u^2\sigma_w^2 - \overline{uw}^2)} b_w^2 [\sigma_u^2 w - \overline{uw}w] + \frac{1}{2}\frac{\partial \sigma_w^2}{\partial z} + \frac{1}{2(\sigma_u^2\sigma_w^2 - \overline{uw}^2)} \times \left[\sigma_w^2 \frac{\partial \overline{wu}}{\partial z} wu - \overline{wu}\frac{\partial \overline{wu}}{\partial z} w^2 - \overline{wu}\frac{\partial \sigma_w^2}{\partial z} wu - \sigma_u^2 \frac{\partial \sigma_w^2}{\partial z} w^2\right]. \tag{11}$$

2.1. Model Parameterization

Within vegetation, U was computed from an exponential relation, first proposed by Cionco (1965),

$$U(z) = U(h) \exp\left(-\alpha\left(1 - \frac{z}{h}\right)\right), \tag{12}$$

where h is canopy height and α is the canopy wind velocity attenuation coefficient. The attenuation coefficient typically ranges between 0.5 and 5, and tends to increase as canopy density progresses from sparse to dense (Cionco, 1978). The logarithmic wind law was applied to calculate horizontal wind velocity above the vegetation:

$$U = \frac{u_*}{k} \ln\left(\frac{z-d}{z_0}\right) \tag{13}$$

where u_* is friction velocity, k is von Karman's constant (0.4), d is the zero plane displacement height and z_0 is the roughness length. The reader should note that Equation (13) applies only for conditions of neutral thermal stratification.

Within vegetation, σ_w/u_* and σ_u/u_* were approximated to decrease linearly with depth (Raupach, 1988) using,

$$\sigma_{w,u}/u_*(z) = a_0 + (a_1 - a_0)z/h. \tag{14}$$

The coefficient, a_1, was assumed to equal 1.25 for σ_w/u_* and 2.39 for σ_u/u_*; the coefficient, a_0, ranges between 0.1 and 0.5 (Raupach, 1988). The vertical variation in the covariance between w and u ($\overline{wu}$) was prescribed to decrease exponentially with depth into the canopy (Uchijima and Wright, 1964).

For most circumstances the Lagrangian time scale (T_L) was assumed to be invariant with height within vegetation (Raupach, 1988, 1989), and was approximated as,

$$T_L = 0.3h/u_*. \tag{15}$$

However, there are data in the literature showing $T_L u_*/h$ to be as low as 0.1 (Amiro, 1990). In the surface boundary layer T_L was approximated, under near neutral stability, as (see Raupach, 1989),

$$T_L = k(z - d)/(1.25u_*). \tag{16}$$

For the theoretical computations reported in this paper, the canopy was considered to be horizontally extensive and homogeneous and the atmosphere was assumed to have a neutral thermal stratification. Flux footprints were computed by releasing a large ensemble of fluid elements ($>$5000) from the forest floor (the algorithms for parcel movement were coded in C language compiled with a 32-bit compiler, and run on a Pentium-class personal computer). The trajectory of each fluid element was tracked and the number of elements that crossed specified heights, after a given travel distance, were counted. The sum of fluid elements captured, at each bin, was normalized by the source strength. Normalization ensured that the integral, with respect to distance (x) at each level, equaled one when integrated between zero and infinity.

Several boundary conditions were specified for the random walk calculations. The time step of each incremental movement was equal to 0.025 $T_L(h)$. Fluid elements reaching the forest floor were reflected perfectly, an *ad hoc* approach that is valid for Gaussian turbulence (Sawford, 1985; Wilson and Sawford, 1996). With regard to wind profile calculations, the zero plane displacement (d) was set at 0.6 h and the roughness parameter (z_0) was 0.1 h. Friction velocity was assumed to equal 0.5 m s^{-1}.

Preliminary tests to debug the model were made by comparing numerical calculations with output from the analytical model of Schuepp et al. (1990). For model calculations representative of the surface boundary layer, both models yielded identical horizontal positions for the peak of the source probability distribution. The numerical model, however, computed a narrower footprint source area than the analytical model. This difference is identical to results published in an earlier analysis by Horst and Weil (1992).

3. Results

Our first objective is to characterize the statistical properties of a ‘flux footprint’ within and above a generic forest canopy for turbulent transfer in the horizontal and vertical dimensions. Figure 1 shows the probability that sources, released at numerous positions upwind, are received downwind at given heights (model parameters are listed in Table I). The most striking feature in Figure 1 is how compact the ‘flux footprint’ source probability function is at various levels within the forest, as compared with the footprint computed for flow above the canopy. At reference heights of 1 and 2 m above the ground, which are typical heights for forest floor flux measurements, the source probability density functions peak at 1 and 3 m

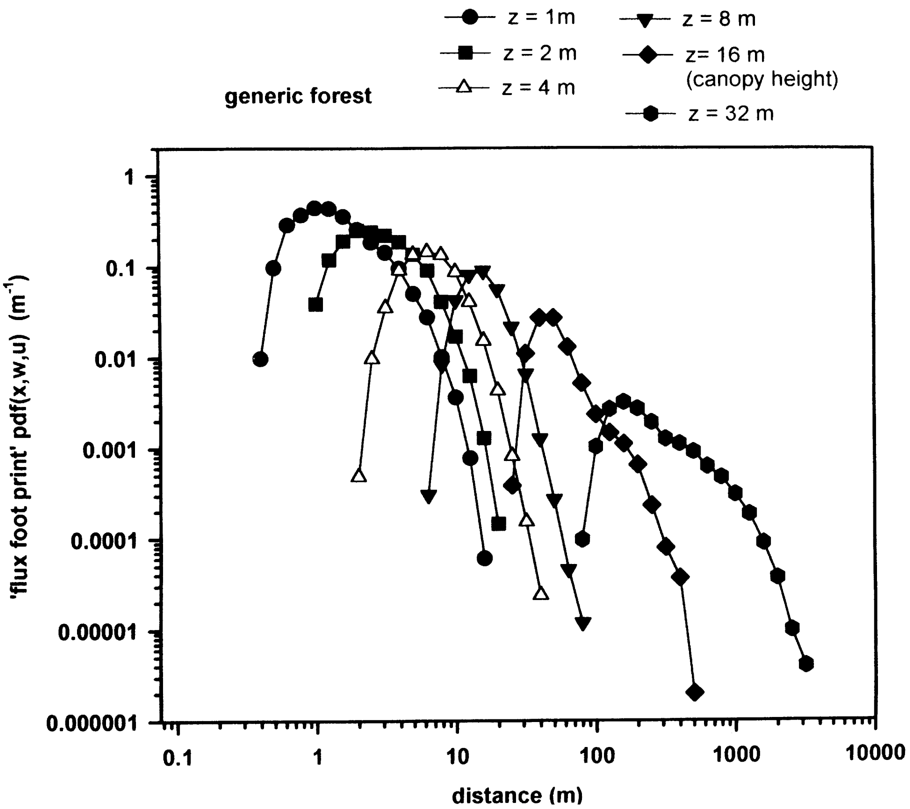

Figure 1. The horizontal distribution of 'flux footprint' probability density functions for arbitrary levels within and above a generic forest. The model calculations considered turbulent fluctuations in the horizontal and vertical dimensions, using the model of Flesch and Wilson (1992). The canopy was 16 m high, $\sigma_w(0)$ was 0.25 u_* and wind attenuation coefficient (α) was 2.5; near-neutral atmospheric stability was assumed. The zero plane displacement (*d*) was 0.60 h and roughness length (z_0) was 0.10 h.

respectively, upwind of the receptor. Furthermore, very little flux information is received from distances beyond 20 m, with regard to these two reference heights. The theoretical size of the flux footprint under a forest is much smaller than the 100 to 1 'rule-of-thumb', fetch-to-height ratio that is often applied for measurements in the surface boundary layer (Gash, 1986).

Figure 1 also shows that the extent of the 'flux footprint', and the position where the probability density function peak, increases with height. At the canopy-atmosphere interface, the shape of the 'flux footprint' probability density function is not mono-modal, but instead possesses an additional bend. This distinctive shape reflects the different movement of fluid parcels within and above the canopy. At twice canopy height (32 m), significant contributions to observed fluxes extend out to about 3 km beyond the reference position, but the 'flux footprint' probability density function peaks at distances between 100 and 200 m upwind of the reference

Table I

Turbulence statistics used to compute 'flux footprint' probability distributions under a generic forest, a boreal conifer forest and a temperate broad-leaved forest. Data on the generic forest stem from the review of Raupach (1988); data on the boreal forest come from Amiro (1990); turbulence data for the temperate forest are from Baldocchi and Meyers (1988)

Variable	Generic forest	Boreal forest	Temperate forest
h (m)	16	14	25
$T_L\ u_*/h$	0.3	0.1	0.1
$\sigma_w(0)/u_*$	0.25	0.325	0.1
$\sigma_w(h)/u_*$	1.25	1.25	1.25
$\sigma_u(h)/u_*$	2.39	2.39	2.35
α	2.5	2.6	3.0
d/h	0.6	0.6	0.85
z_0/h	0.1	0.1	0.1

tower. Though rapid, this theoretical adjustment of the internal boundary layer is consistent with information reported by Leclerc and Thurtell (1990) and Schuepp et al. (1990) for aerodynamically rough canopies.

Figure 2 compares source probability density functions, with respect to distance, for situations when horizontal displacements of fluid parcels are, and are not, a function of fluctuating horizontal wind velocity. Inside the canopy, at 2 m above the ground (Figure 2a), the probability density functions are narrower, when horizontal and vertical wind fluctuations are considered, than situations when only w fluctuations are considered, though the positions of the peaks of the probability distributions remain similar. For the meteorological conditions considered, the 'flux footprint' probability density function, derived from the two-dimensional model, extends out to only 20 m, while the 'flux footprint', derived from the one-dimensional model, extends out to 65 m. Hence, use of the simpler one-dimensional model, reported in Thomson (1987), instead of the two-dimensional version, reported in Flesch and Wilson (1992), would cause one to draw an erroneous conclusion about the horizontal dimensions of the 'flux footprint' inside a forest. With respect to flow near the canopy-atmosphere interface and in the surface boundary layer (Figures 2b and 2c), the characteristics of the 'flux footprint' probability density distribution are not markedly altered. This conclusion is reasonable, since turbulence intensities are rather small ($<$20%) in those regions.

The intensity of vertical turbulent mixing deep in the canopy has a theoretical potential to impact upon the statistical characteristics of 'flux footprints' within a forest. Conceptually, circumstances that experience large σ_w enable fluid elements to reach a given reference height sooner than conditions that maintain low values of σ_w. Figure 3 shows that the horizontal location of the footprint probability peak increases four-fold, at a reference height of 2 m, as σ_w/u_* near the forest floor,

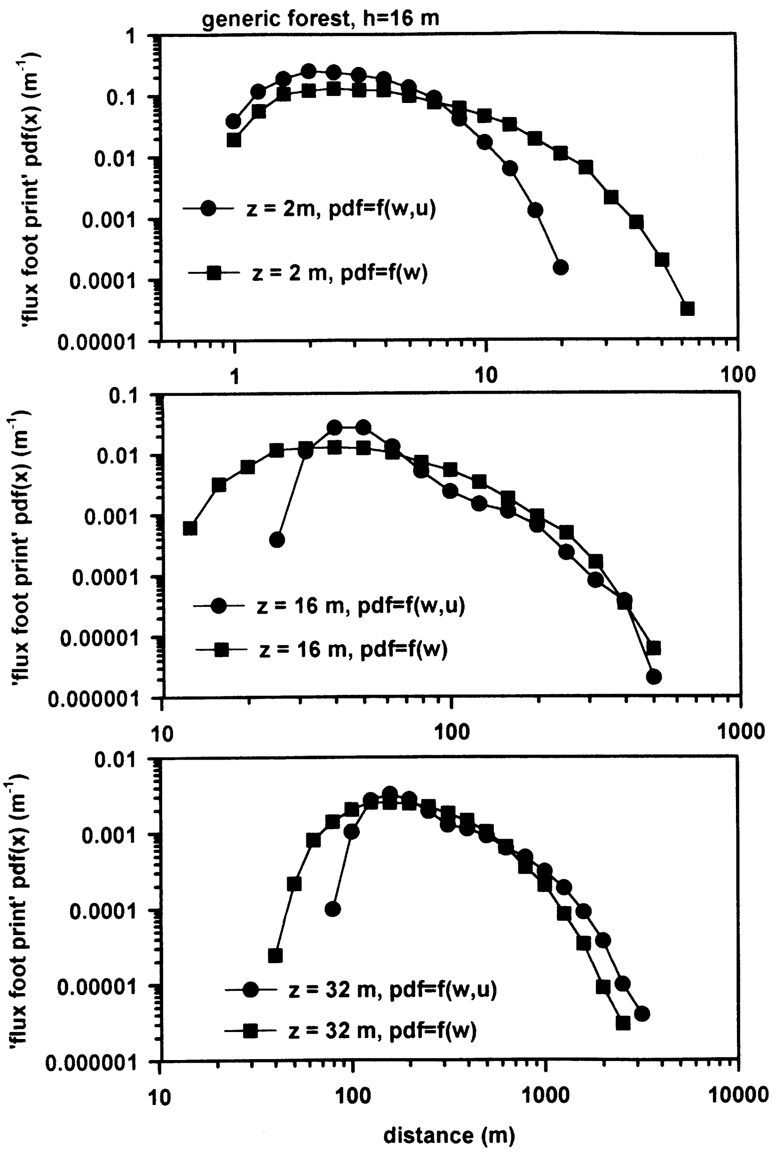

Figure 2. The horizontal distribution of 'flux footprint' probability density functions for arbitrary levels within and above a generic forest. In one case, horizontal movement of fluid parcels is a function of the mean and fluctuating wind velocity. The second case forces horizontal movement of fluid parcels to be a function of the mean wind velocity only.

decreases from 0.4 to 0.1. Significant contributions from sources further upwind are also observed when vertical turbulent mixing is relatively low, though this effect is not large relatively.

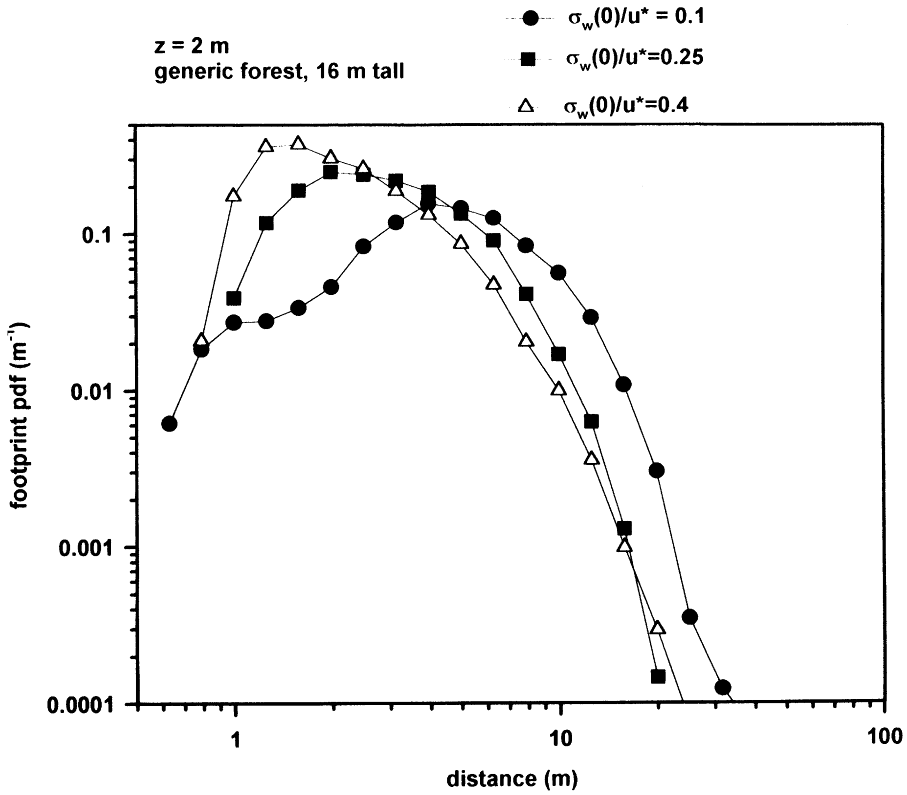

Figure 3. The horizontal distribution of 'flux footprint' probability density functions assessed theoretically at 2 m above the floor of a generic forest. These calculations are a function of the standard deviation of vertical velocity fluctuations measured near the forest floor ($\sigma_w(0)$). The canopy height was 16 m and α was 2.5.

How fast a fluid element moves horizontally should also impact upon the horizontal spread of the flux footprint and the location of its peak. With this in mind, the effect of the within-canopy wind attenuation coefficient (α) on 'flux footprint' probability density distributions is examined. Sparse open vegetation, such as savanna and woodlands, possess values of α close to one (Cionco, 1965, 1978). For this case, the detected flux footprint source area, at a height of 2 m, extends horizontally out to 60 m and its probability density function peaks at about 7 m upwind of the reference position (Figure 4). For the case representing a very dense canopy, such as temperate and tropical broad-leaved forests, α was assigned a value of four. Under this extreme condition, the 'flux footprint', at a reference height of 2 m, is confined within 10 m and its probability density function peaks at about 0.8 m. The intermediate case (α equals 2.5) is representative of wind flow within many crops and boreal forests (Cionco, 1978; Amiro, 1990). Here the flux probability density function peaks at 2 m and the footprint source area extends out to about 20 m.

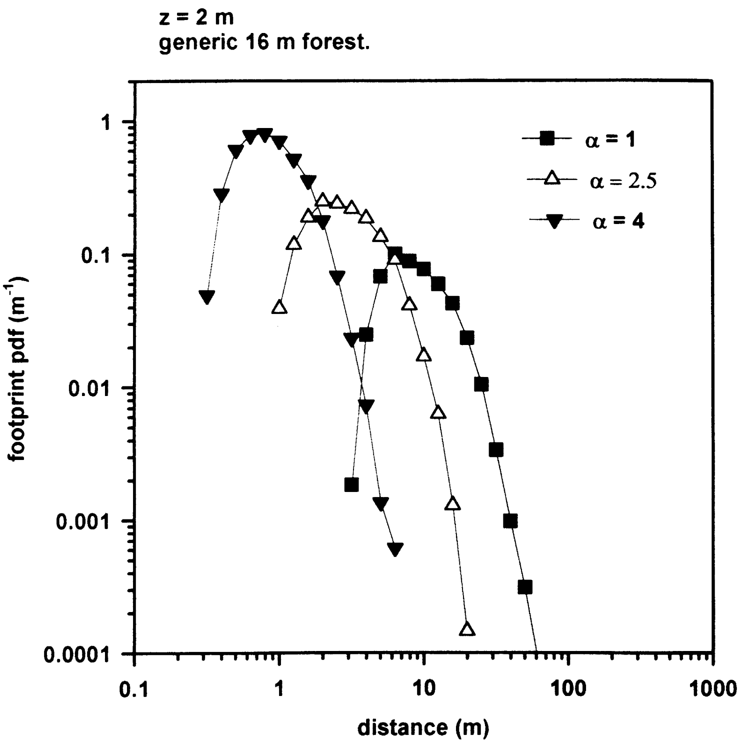

Figure 4. The impact of the horizontal wind speed attenuation coefficient (α) on the horizontal distribution of 'flux footprint' probability density functions assessed at 2 m above the floor of a generic forest. The canopy height was 16 m and $\sigma_w(0)$ was 0.25 u_*.

Wind speed and σ_w are relative functions of canopy height, so their absolute values and the statistical characteristics of the 'flux footprint' at a given level are expected to vary with canopy height. Figure 5 shows that canopy height has a weak impact on the position of the peak of the 'flux footprint', at 2 m, but the extent of the flux footprint extends length-wise markedly under taller forests. Significant contributions to the probability density function reach beyond 100 m under a 30 m tall forest.

Although the Markovian movement of fluid parcels is a function of the Lagrangian time scale, 'flux footprint' probability distribution calculations were rather insensitive to a three-fold variation in the parameter defining the Lagrangian time scale (Figure 6).

In nature, interactions between turbulence and canopy structure cause unique combinations of turbulence model parameters to occur. To account for this situation we applied the footprint model to two field cases. One case involves a sparse, 14-m tall boreal conifer forest (Figure 7) and the other a 25-m tall dense, temperate broad-leaved forest (Figure 8) (see Table I for model parameters). Figure 7 shows that the flux footprint probability density function at 2 m above the floor of the

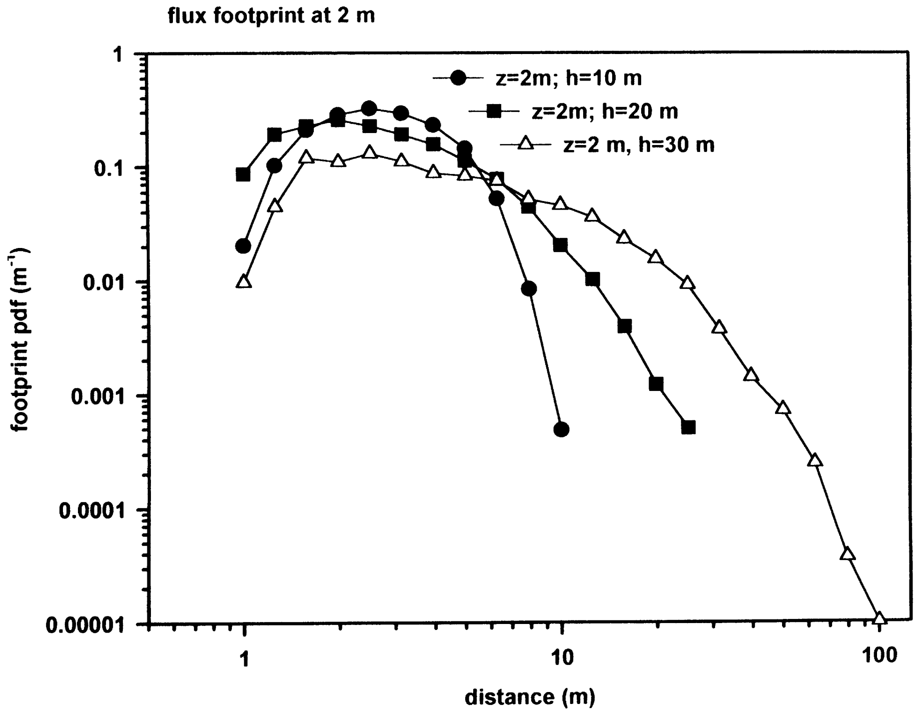

Figure 5. The impact of canopy height on the horizontal distribution of 'flux footprint' probability density functions assessed at 2 m above the floor of a generic forest; $\sigma_w(0)$ was 0.25 u_* and α was 2.5.

boreal forest peaks 2 m upwind from the tower and the 'flux footprint' extends out to 10 m. For eddy fluxes measured above the boreal forest canopy, at 32 m, the source probability density function peaks, theoretically, at 200 m and extends out to 10,000 m. In contrast, the 'flux footprint' for the temperate deciduous forest (Figure 8) is contracted; the 'flux footprint' at 2 m peaks at 3 m, and is restricted within 30 m of the tower. The above canopy 'flux footprint' peaks at 125 m and extends 2,000 m upwind, a much different behaviour than the one noted for the boreal forest (Figure 7).

Anecdotal evidence supports these calculations of small, flux footprint source areas under forests. Whenever the author approached his flux measurement systems located under a boreal or temperate forest, he only saw spikes on the CO_2 signal from his breath when he was within 5 m of the tower. He never saw the influence of colleagues walking beyond 30 m of the instrument arrays.

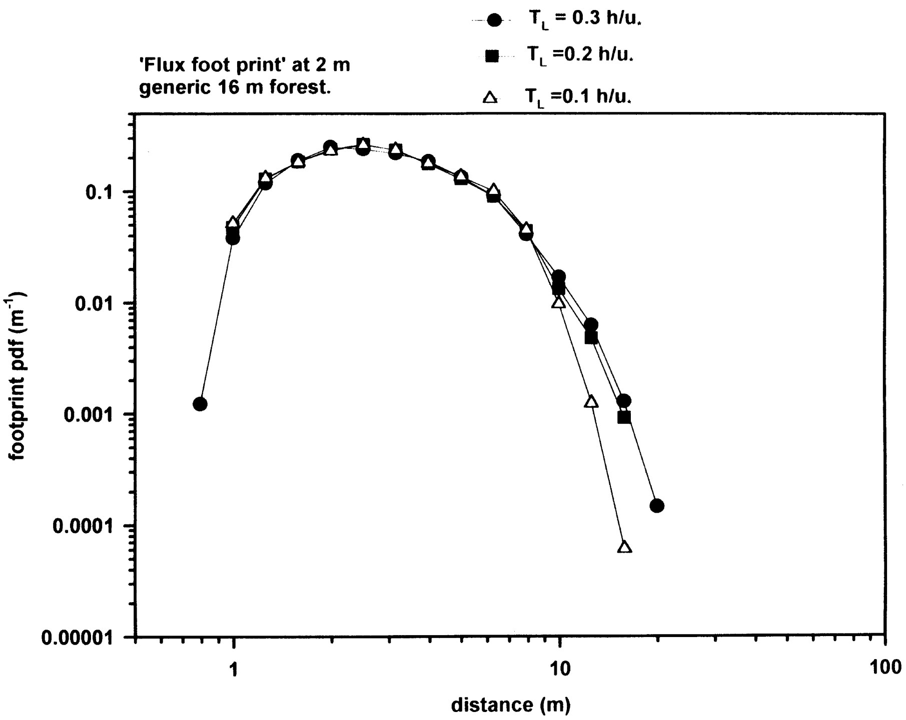

Figure 6. The impact of the Lagrangian time scale on the horizontal distribution of 'flux footprint' probability density functions assessed at 2 m above the floor of a generic forest. The canopy was 16 m tall, $\sigma_w(0)$ was 0.25 u_* and α was 2.5.

4. Discussion

An attempt to synthesize the impact of U and σ_w on characteristics of flux footprints is made by using non-dimensional variables. Figure 9 shows that the horizontal distance where the flux footprint probability density function peaks ($x_{\max}$), relative to the reference height (z), increases with respect to the ratio between U and σ_w. Adopting a linear fit through the data yields the conclusion that doubling of U/σ_w doubles the position of the normalized footprint peak. These results support (and quantify) the contention that flux footprints extend the farthest under canopies that are aerodynamically transparent, or canopies subjected to conditions that restrict vertical fluctuations of air motion.

The lower portion of canopies tends to possess a stable thermal stratification by day and an unstable one by night due to the interception of radiation in the crown (Jacobs et al., 1996). The impact of thermal stability on within-canopy flux footprints was not evaluated directly because there is no established theory that characterizes the impact of thermal stability on turbulence statistics within a canopy. Nevertheless, the consequences of different stability conditions can be inferred

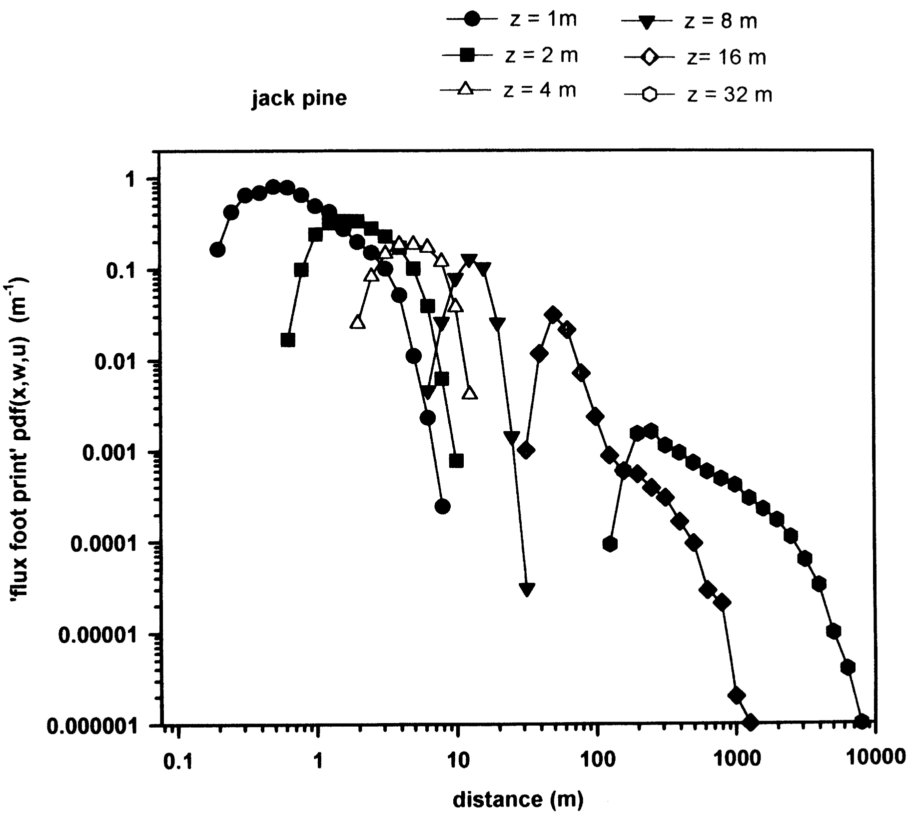

Figure 7. The horizontal distribution of 'flux footprint' probability density functions below and above a boreal jack pine forest. The Lagrangian random walk model was parameterized with turbulence statistics reported by Amiro (1990) (see Table I).

through the examination of how changes in σ_w and U affected characteristics of the flux footprint. In concept, stable stratification will act to reduce σ_w, and it increases U relative to u_* (e.g., Raupach et al., 1996; Baldocchi and Meyers, 1988). Together, these attributes will extend the flux footprint under stable conditions, as is reported by Leclerc and Thurtell (1990) for the surface boundary layer.

The data presented in Figures 7 and 8 encourage a re-interpretation of our sub-canopy flux measurements below these two forests (see Baldocchi and Vogel, 1996). For the temperate forest study, solar radiation measurements were made along a 30-m traverse that was within 30 m of the flux measurement site. Consequently, radiation measurements were made near the flux footprint sensed by the eddy covariance system. In this case, measurements of net radiation and the sum of independently measured energy balance components (soil, latent and sensible flux densities) agreed within 1%, on average. In the case of the boreal forest study, solar radiation measurements were made along a 16 m transect that was placed more than 70 m away from the sub-canopy flux tower. In retrospect, that area

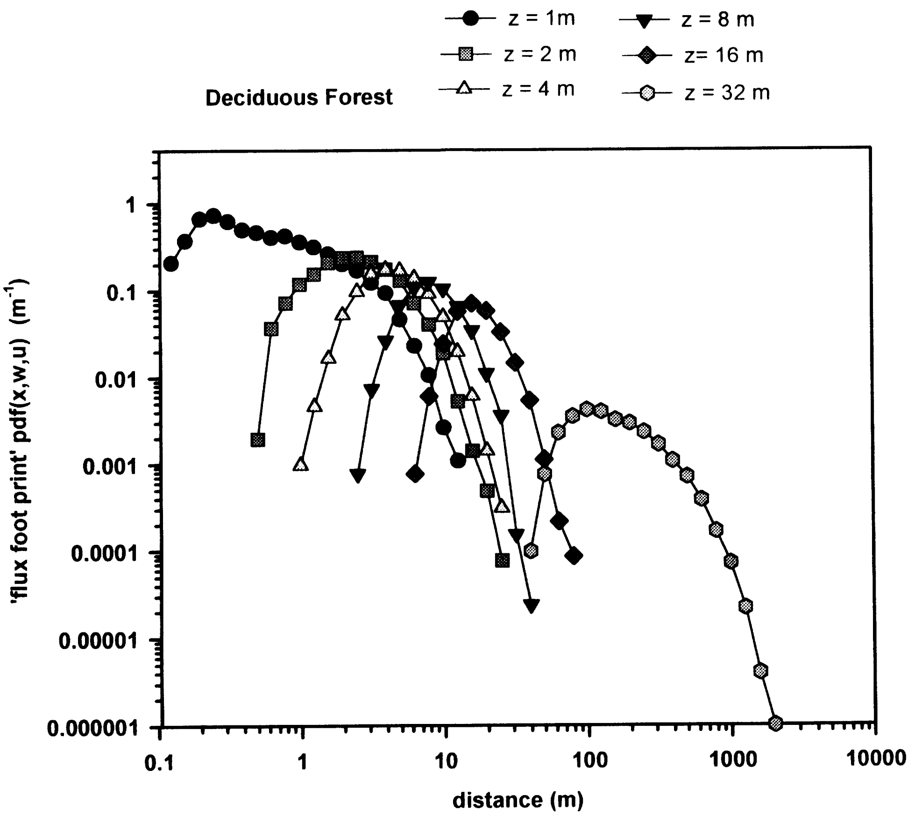

Figure 8. The horizontal distribution of 'flux footprint' probability density functions below and above a temperate deciduous forest. The Lagrangian random walk model was parameterized with turbulence statistics reported by Baldocchi and Meyers (1988) (see Table I).

possessed a higher leaf area index than the area in the vicinity of the flux tower (Chen, 1996). Consequently, more solar radiation penetrated the canopy near the flux measurement system. This effect probably contributed to the observation that the measured sums of latent, sensible and soil heat flux exceeded the measurement of net radiation significantly (see Baldocchi and Vogel, 1996).

5. Summary and Conclusions

A Lagrangian random walk model was applied to the task of computing 'flux footprint' probability density functions below and above a forest canopy. The study is warranted because the turbulence regime beneath vegetation is inhomogeneous and distinct from that in the surface boundary layer (Raupach, 1988), where 'foot print' calculations have been done in the past (Leclerc and Thurtell, 1990; Horst and Weil, 1990). The source area of 'flux footprints' detected by a tower under a

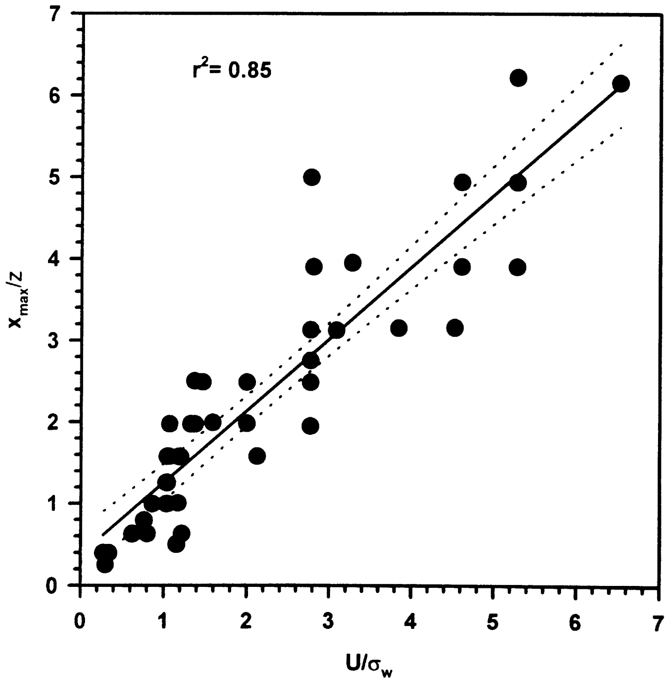

Figure 9. The relationship between the distance of the peaks of the 'flux footprint' probability density function (normalized by the measurement height) and the ratio between horizontal wind velocity (U) and $\sigma_w(z)$.

canopy is much contracted, as compared to one evaluated in the surface boundary layer. In general, the 'flux footprint' probability function for a tower placed 2 m above a forest floor, peaks within 5 m of the tower and its extent is restricted within 40 m. It should also be stressed that intepretation of the flux footprint, derived from a model that only considers vertical velocity fluctuations, differs from that derived from a more, realistic, two-dimensional model.

With respect to this analysis, future experimental work is required to verify these calculations. One option would involve measuring evaporation below an orchard or forest, whose soil is covered with evaporating grass. The extent of flux footprints could be tested by measuring fluxes over an array of treatments that artificially cover segments of the grass with a plastic film. Verification of fluxes could be achieved by comparing eddy covariance measurements with estimates derived from a network of mini-lysimeters (Black and Kelliher, 1989). Visualization of the advection and diffusion of a source underneath a canopy could be conducted with the combined use of smoke releases and video imaging.

Wilson and Sawford (1996) identified important and unresolved issues relating to the quantification of a non-Gaussian random forcing term and non-perfect reflection of fluid elements under a forest with non-Gaussian turbulence. The validity of applying a Lagrangian model, with Gaussian forcing, to assess 'flux footprint' characteristics in a situation with non-neutral thermal stratification and non-Gaussian turbulence deserves additional investigation.

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References

Amiro, B. D.: 1990, 'Comparison of Turbulence Statistics Within Three Boreal Forest Canopies', *Boundary-Layer Meteorol.* **51**, 99–121.

Baldocchi, D. D., Verma, S. B., and Anderson, D. E.: 1987, 'Canopy Photosynthesis and Water Use Efficiency in a Deciduous Forest', *J. Appl. Ecol.* **24**, 251–260.

Baldocchi, D. D. and Meyers, T. P.: 1988, 'Turbulence Structure in a Deciduous Forest', *Boundary-Layer Meteorol.* **43**, 345–365.

Baldocchi, D. D. and Meyers, T. P.: 1991, 'Trace Gas Exchange at the Floor of a Deciduous Forest: I. Evaporation and CO_2 Efflux', *J. Geophys. Res. (Atmos.)* **96**, 7271–7285.

Baldocchi, D. D. and Vogel, C.: 1996, 'A Comparative Study of Water Vapor, Energy and CO_2 Flux Densities Above and Below a Temperate Broadleaf and a Boreal Pine Forest', *Tree Physiol.* **16**, 5–16.

Biscoe, P. V., Scott, R. K., and Monteith, J. L.: 1975, 'Barley and its Environment. III. Carbon Budget of the Stand', *J. Appl. Ecol.* **12**, 269–293.

Black, T. A. and Kelliher, F. M.: 1989, 'Processes Controlling Understory Evaporation', *Phil. Trans. Roy. Soc. Lond. B.* **324**, 207–231.

Black, T. A., den Hartog, G., Neumann, H., Blanken, P., Yang, P., Nesic, Z., Chen, S., Russel, C., Voroney, P., and Staebler, R.: 1996, 'Annual Cycles of CO_2 and Water Vapour Fluxes Above and Within a Boreal Aspen Stand', *Global Change Biol.* **2**, 219–230.

Chen, J.: 1996, 'Optically-Based Methods for Measuring Seasonal Variation of Leaf Area Index in Boreal Conifer Stands', *Agric. For. Meteorol.* **80**, 135–163.

Cionco, R. M.: 1965, 'A Mathematical Model for Air Flow in a Vegetative Canopy', *J. Appl. Meteorol.* **4**, 517.

Cionco, R. M.: 1978, 'Analysis of Canopy Index Values for Various Canopy Densities', *Boundary-Layer Meteorol.* **15**, 81–93.

Durbin, P. A.: 1983, *Stochastic Differential Equations and Turbulent Dispersion*, NASA Reference Publication 1103.

Finn, D., Lamb, B., Leclerc, M. Y., and Horst, T. W.: 1996, 'Experimental Evaluation of Analytical and Lagrangian Surface-Layer Flux Footprint Models', *Boundary-Layer Meteorol.* **80**, 283–308.

Finnigan, J. J. and Brunet, Y.: 1995, 'Turbulent Airflow in Forests on Flat and Hilly Terrain', in M. P. Coutts and J. Grace (eds.), *Wind and Trees*, Cambridge University Press, Cambridge, pp. 3–40.

Flesch, T. K.: 1996, 'The Footprint for Flux Measurements, From Backward Lagrangian Stochasitic Models', *Boundary-Layer Meteorol.* **78**, 399–404.
Flesch, T. K. and Wilson, J. D.: 1992, 'A Two Dimensional Trajectory-Simulation Model for Non-Gaussian, Inhomogeneous Turbulence Within Plant Canopies', *Boundary-Layer Meteorol.* **61**, 349–374.
Freijer, J. I. and Bouten, W.: 1991, 'A Comparison of Field Methods for Measuring Soil Carbon Dioxide Evolution: Experiments and Simulation', *Plants and Soil* **135**, 133–142.
Gash, J. H. C.: 1986, A Note on Estimating the Effect of a Limited Fetch on Micrometeorological Evaporation Measurements', *Boundary-Layer Meteorol.* **35**, 409–415.
Hollinger, D. Y., Kelliher, F. M., Byers, J. N., Hunt, J. E., McSeveny, T. M., and Weir, P. L.: 1994, 'Carbon Dioxide Exchange Between an Undisturbed Old-Growth Temperate Forest and the Atmosphere', *Ecology* **75**, 134–150.
Horst, T. W. and Weil, J. C.: 1992, 'Footprint Estimation for Scalar Flux Measurements in the Atmospheric Surface Layer', *Boundary-Layer Meteorol.* **59**, 279–296.
Hutchinson, G. L. and Livingston, G. P.: 1993, 'Use of Chamber Systems to Measure Trace Gas Fluxes', in L. A. Harper et al. (eds.), *Agricultural Ecosystem Effects on Trace Gases and Global Climate Change*, American Society of Agronomy, Madison, WI, pp. 63–78.
Jacob, D. and Bakwin, P. S.: 1991, 'Cycling of NO_x in Tropical Forest Canopies', in J. Rogers and W. Whitman (eds.), *Microbial Production and Consumption of Greenhouse Gases, Am. Soc. Microbiol.* Washington, DC, pp. 237–254.
Jacobs, A. F. G., van Boxel, J. H., and El-Kilani, R. M. M.: 1994, 'Nighttime Free Convection Characteristics Within a Plant Canopy', *Boundary-Layer Meteorol.* **71**, 375–391.
Kaimal, J. C. and Finnigan, J. J.: 1994, *Atmospheric Boundary Layer Flows*, Oxford University Press, Oxford, 289 pp.
Lavigne, M. B., Ryan, M. G., Anderson, D. E., Baldocchi, D. D., Crill, P. M., Fitzjarrald, D. R., Goulden, M. L., Gower, S. T., Massheder, J. M., McCaughey, J. H., Rayment, M., and Striegl, R. G.: 1997, 'Comparing Nocturnal Eddy Covariance Measurements to Estimates of Ecosystem Respiration Made by Scaling Chamber Measurements', *J. Geophys. Res.*, in press.
Leclerc, M. Y. and Thurtell, G. W.: 1990, 'Footprint Predictions of Scalar Fluxes Using a Markovian Analysis', *Boundary-Layer Meteorol.* **52**, 247–258.
Leclerc, M. Y., Shen, S., and Lamb, B.: 1997, 'Observations and Large-Eddy Simulation Modeling of Footprints in the Lower Convective Boundary Layer', *J. Geophys. Res.* **102**, 9323–9334.
Luhar, A. K. and Britter, R.: 1989, 'Random Walk Model Studies for Dispersion in Inhomogeneous Turbulence in Convective Boundary Layers', *Atmos. Environ.* **23**, 1911.
Luhar, A. K. and Rao, K. S.: 1994, 'Source Footprint Analysis for Scalar Fluxes Measured in Flows Over Inhomogeneous Surface', in S. V. Gryning and M. M. Millan (eds.), *Air Pollution Modelling and Its Application*, Plenum Press, New York, pp. 315–323.
Raupach, M. R.: 1988, 'Canopy Transport Processes', in W. L. Steffen and O. T. Denmead (eds.), *Flow and Transport in Natural Environment: Advances and Application*, Springer-Verlag, Berlin, pp. 95–127.
Raupach, M. R.: 1989, 'Applying Lagrangian Fluid Mechanics to Infer Scalar Source Distributions from Concentration Profiles in Plant Canopies', *Agric. For. Meteorol.* **47**, 85–108.
Raupach, M. R., Finnigan, J. J., and Brunet, Y.: 1996, 'Coherent Eddies and Turbulence in Vegetation Canopies: The Mixing-Layer Analogy', *Boundary-Layer Meteorol.* **78**, 351–382.
Rochette, P., Desjardins, R. L., Pattey, E., and Lessard, R.: 1996, 'Instantaneous Measurement of Radiation and Water Use Efficiencies of a Maize Crop', *Agronomy J.* **88**, 627–635.
Saugier, B., Granier, A,. Pontailler, J. Y., Dufrene, E., and Baldocchi, D. D.: 1997, 'Transpiration of a Boreal Pine Forest, Measured by Branch Bags, Sapflow and Micrometeorological Methods', *Tree Physiol.* **17**, 511–519.
Sawford, B. L.: 1985, 'Lagrangian Statistical Simulation of Concentration Mean and Fluctuation Fields', *J. Clim. Appl. Meteorol.* **24**, 1152–1166.
Schmid, H. P.: 1994, 'Source Areas for Scalars and Scalar Fluxes', *Boundary-Layer Meteorol.* **67**, 293–318.

Schuepp, P. H., Leclerc, M. Y., MacPherson, J. I., and Desjardins, R. L.: 1990, 'Footprint Prediction of Scalar Fluxes from Analytical Solutions of the Diffusion Equation', *Boundary-Layer Meteorol.* **50**, 355–373.

Thomson, D. J.: 1987, 'Criteria for the Selection of Stochastic Models of Particle Trajectories in Turbulent Flow', *J. Fluid Mech.* **180**, 529–556.

Uchijima, Z. and Wright, J. L.: 1964, 'An Experimental Study of Air Flow in a Corn Plant Air Layer', *Bull. National Inst. Agric. Sci. (Japan)* **11A**, 19–44.

Wilson, J. D.: 1989, 'Turbulent Transport Within the Plant Canopy', in T. A. Black, D. L. Spittlehouse, M. D. Novak, and D. T. Price (eds.), *Estimation of Areal Evapotranspiration*, IAHS Publication 177, Wallingford, U.K., pp. 43–80.

Wilson, J. D. and Sawford, B. L.: 1996, 'Review of Lagrangian Stochasitc Models for Trajectories in the Turbulent Atmosphere', *Boundary-Layer Meteorol.* **78**, 191–210.