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Flux footprint simulation downwind of a forest edge

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Abstract Surface fluxes, originating from forest patches, are commonly calculated from atmospheric flux measurements at some height above that patch using a correction for flux arising from upwind surfaces. Footprint models have been developed to calculate such a correction. These models commonly assume homogeneous turbulence, resulting in a simulated atmospheric flux equal to the average surface flux in the footprint area. However, atmospheric scalar fluxes downwind of a forest edge have been observed to exceed surface fluxes in the footprint area. Variations in atmospheric turbulence downwind of the forest edge, as simulated with an $E - \varepsilon$ model, can explain enhanced atmospheric scalar fluxes. This $E - \varepsilon$ model is used to calculate the footprint of atmospheric measurements downwind of a forest edge. Atmospheric fluxes appear mainly enhanced as a result of a stronger sensitivity to fluxes from the upwind surface. A sensitivity analysis shows that the fetch over forest, necessary to reach equilibrium between atmospheric fluxes and surface fluxes, tends to be longer for scalar fluxes as compared to momentum fluxes. With increasing forest density, atmospheric fluxes deviate even more strongly from surface fluxes, but over shorter fetches. It is concluded that scalar fluxes over forests are commonly affected by inhomogeneous turbulence over large fetches downwind of an edge. It is recommended to take horizontal variations in turbulence into account when the footprint is calculated for atmospheric flux measurements downwind of a forest edge. The spatially integrated footprint is recommended to describe the ratio between the atmospheric flux and the average surface flux in the footprint.

Keywords Footprint · Forest edge · Scalar flux · Simulation · Turbulence

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1 Introduction

Atmospheric flux measurements reflect surface fluxes in an upwind source area; ‘flux’ refers to a vertical turbulent flux of an entity (e.g. heat, CO_2, momentum). The sensitivity of the atmospheric flux to the surface flux at a certain location is given by the footprint function. Conversely, surface fluxes can be calculated from an atmospheric observation above the investigated patch using a footprint-derived correction for surface fluxes from the upwind area. This option makes the footprint especially useful in complex terrain.

Conservation of mass and energy requires that the atmospheric flux equals the average surface flux of the footprint area in the case of stationary, homogeneous turbulence and in the absence of atmospheric sinks or sources. Observations by Klaassen et al. (2002) show, in contrast, enhanced atmospheric heat fluxes downwind of a forest edge; with ‘enhanced’ we mean higher that the average surface flux in the footprint area. Simulations by Sogachev et al. (2004) showed large variations in the atmospheric scalar flux over hilly terrain with a spatially constant surface flux, also implying that the atmospheric flux deviates from the average surface flux in the footprint area. These results indicate that deviations between atmospheric fluxes and average surface fluxes are realistic and should be accounted for in footprint models.

Most footprint models nowadays depend on the assumption of homogeneous turbulence and models depending on this assumption will be denoted ‘common’ footprint models. Turbulence in complex terrain is seldom homogeneous, especially in situations with orography, irregular placed obstacles, or landscapes with large variations in surface roughness. All atmospheric fluxes are related to turbulence, so inhomogeneous turbulence may cause the atmospheric flux to deviate locally from the average surface flux in the footprint. We hypothese that common footprint models can be improved when horizontal variations in turbulence downwind of a forest edge are taken into account. This hypothesis will be validated using an updated version of the model of Sogachev et al. (2004).

2 Method

2.1 Theoretical consideration

The footprint function f (m^{-1}) describes the sensitivity of a vertical turbulent flux η at r to the surface forcing Q at separation (r'). In integral form (Pasquill and Smith 1983; see also Schmid 2002):

$$\eta(r) = \int_R Q(r + r') f(r, r')\, dr' \tag{1}$$

where R is the integration domain. In order to evaluate the impact of inhomogeneous turbulence on atmospheric flux measurements we adopt the integrated footprint function, derived from Eq. (1) as

$$F_R = \int_R f(r, r')\, dr'. \tag{2}$$

The integrated footprint describes the ratio between atmospheric fluxes and average surface fluxes in the footprint area, so F_R shows the weakening or strengthening of atmospheric fluxes as compared to average surface fluxes. Conservation laws require that the integrated footprint reaches unity ($F_R = 1$) in stationary conditions with horizontal homogeneity of both flow and turbulence (Haenel and Grünhage 1999; Kormann and Meixner 2001). Differences between average surface fluxes in the footprint area and atmospheric fluxes ($F_R \neq 1$) may arise when the measurement height is no longer small compared to the boundary-layer height (Horst and Weil 1994) due to the entrainment fluxes higher up in the mixed layer, or locally induced by inhomogeneous turbulence. The present study is focussed on the impact of inhomogeneous turbulence on atmospheric flux measurements.

The land surface is often assumed to consist of a mosaic of patches (Koster and Suarez 1992) with uniform surface fluxes within a patch and different surface fluxes between patches. Here we will assume a simple landscape consisting of a forest patch and a single upwind patch, so $F_R = F_\mathrm{i} + F_\mathrm{u}$ with F_i the integrated footprint over the investigated patch and F_u the integrated footprint over the upwind patch. Uniform surface forcing in the patches results in

$$F_\mathrm{b} = \int_0^{r''} f(r'|z_m)dr', \tag{3a}$$

and

$$F_\mathrm{u} = \int_{r''}^{\infty} f(r'|z_m)\,dr' \tag{3b}$$

where r'' is the edge of the patch (Gash 1986; Horst and Weil 1994). Using Eq. (1), the atmospheric flux over the forest, with surface fluxes Q_i and Q_u for investigated and upwind patches respectively, is then given as

$$\eta(0,\ z_\mathrm{m}) = Q_\mathrm{i}F_\mathrm{i} + Q_\mathrm{u}F_\mathrm{u} = F_RQ_\mathrm{i} + F_\mathrm{u}(Q_\mathrm{u} - Q_\mathrm{i}). \tag{4}$$

Thus, the surface flux of the patch of investigation is given by

$$Q_\mathrm{i} = \frac{\eta(0,\ z_\mathrm{m}) - F_\mathrm{u}Q_\mathrm{u}}{F_R - F_\mathrm{u}}. \tag{5}$$

Variations in turbulence result in additional terms in the atmospheric conservation equations. By assuming incompressible flow and neglecting the Leonard terms (Leonard 1974), mass conservation results in a relation between the atmospheric scalar flux $\eta(0, z_\mathrm{m})$ and the flux of the underlying surface Q_i :

$$\eta(0, z_\mathrm{m}) = \overline{(w'c')} = Q_\mathrm{i} - \int_0^{z_m} \overline{u}\,\frac{\partial c}{\partial x}dz - \int_0^{z_m} \overline{w}\frac{\partial c}{\partial z}dz \tag{6}$$

where symbols with overbars and primes denote respectively mean and fluctuating values for horizontal (u) and vertical (w) velocity components, where c is the scalar. The second term of the right-hand side of Eq. (6) describes horizontal advection and can be calculated from upwind fluxes using common footprint models. The last term of (6) describes vertical advection. Homogeneous flow implies $\overline{w} = 0$ (mass conservation), no vertical advection, and $F_R = 1$. Thus, inhomogeneous flow is characterised by inclusion of vertical advection and vertical advection causes the turbulent flux to deviate from the average surface flux in the footprint, so $F_R \neq 1$. As a result, F_R is in theory a useful parameter to characterise vertical advection in the context of footprint modelling.

2.2 Model description

The theoretical consideration has shown that horizontal as well as vertical advection influence the relation between atmospheric fluxes and surface fluxes of the patch under investigation. Both advective fluxes are calculated here using a single model of atmospheric turbulence.

An updated version of the model of Sogachev et al. (2005) is used as this model has already been validated for footprint calculations in hilly terrain covered with forest. The model is two-dimensional and assumes a neutrally stratified surface layer. In the present study the model is used to calculate the footprint in a partially forested flat landscape. Atmospheric flow is calculated using $E - \varepsilon$ closure, with E is the turbulent kinetic energy (TKE) and ε is the dissipation rate of TKE. Since details about numerical schemes that solve the system of non-linear differential equations (mixed parabolic-elliptic type in the two-dimensional case) can be found in Sogachev et al. (2005), we provide here only the information that has not been discussed previously, or is essential for the present investigation.

Unlike Sogachev et al. (2005) we use a new parameterisation of additional source/sink terms for E and ε within the canopy, based on an update of the model coefficient C_2 determining the rate of turbulence decay within a vegetation canopy, given by Sogachev and Lloyd (2004):

$$C_2^* = C_2 - \frac{(C_2 - C_1)\, S_{\mathrm{d}}}{\varepsilon}, \tag{7}$$

where C_1 is the coefficient by shear production term and S_{d} denotes dissipation due to plant drag. The latter is expressed as (Sanz 2003)

$$S_{\mathrm{d}} = \beta_{\mathrm{d}} c_{\mathrm{d}} A(z) U E, \tag{8}$$

where U is the mean velocity of the air flow, $A(z)$ is the leaf area per unit volume of space, c_{d} is the drag coefficient for unit plant area density, taken as $c_{\mathrm{d}} = 0.2$, and β_{d} is the TKE loss ratio on interactions with obstacles. Coefficient β_{d} varies in different studies between 2 and 4, and here a value of 3.33 is used. The leaf area density $A(z)$ is calculated from Markkanen et al. (2002):

$$A\left(L_0, \frac{z}{h}, \alpha\right) = L_0 \frac{(z/h)^{\alpha-1}\,(1 - z/h)^2}{\int\limits_0^1 (z/h)^{\alpha-1}\,(1 - z/h)^2\, d(z/h)} \tag{9}$$

where h is the canopy height, L_{o} is leaf area index ($\mathrm{m}^2\ \mathrm{m}^{-2}$) and α is a parameter describing the shape of the distribution; with increasing α more leaves are located near the top of the vegetation. The denominator of (9) is included to normalise the function.

The vertical distribution of forcing $Q(z)$ of a scalar c inside the forest is calculated using

$$Q\left(z\right)=\gamma A(z)\exp\left(-\frac{0.5L_0(z)}{\sin h_s}\right)c_1 \tag{10}$$

where $L_o(z)$ is a cumulative leaf area index, h_s is the sun elevation, taken as $h_s = 45°$, c_1 denotes scalar concentration on the leaf surface and γ ($\mathrm{m\,s^{-1}}$) is a constant of proportionality fitted to simulate the observed fluxes above bog and forest. The format of Eq. (10) is selected to describe the extinction of unidirectional radiation inside a canopy with randomly oriented leaves. As such, Eq. (10) is expected to be representative of the source of energy fluxes within the forest canopy. The scalar forcing at the soil surface inside the forest is set to zero. Flux footprint and integrated footprint along the bog–forest transition have been calculated from Eq. (1) using constant surface forcing Q over the full integration domain. A closed upper boundary without turbulent fluxes is assumed in the model resulting in a linear decrease of fluxes with height over a homogeneous surface. Such a decrease of fluxes is common in the atmospheric boundary layer during daytime. Conservation of mass implies that vertical flux divergence causes a slow temporal change of scalar concentrations in the boundary layer, described as ‘quasi-stationary’. In the calculations, the surface forcing is selected in such a way that a homogenous surface would result in unit atmospheric flux at measurement level, so deviations from unity at measurement level reflect on the influence of surface heterogeneity and vertical advection.

2.3 Model initialisation

Numerical experiments have been carried out inside the computational space (8,500 m and 3,025 m in the horizontal and vertical directions, respectively) divided into cells using a 84 × 103 grid. To minimise the effect of lateral boundary conditions, a variable model horizontal step was used changing from 500 m at the borders to 25 m in the central zone around the forest edge. A vertical irregular grid (with a minimum step at the surface of 0.17 m and a maximum at the upper boundary of 200 m) provided the necessary accuracy for a detailed description of the canopy structure and the exchange processes occurring within it. Surface roughness and energy fluxes for the upwind bog were taken constant over the domain, with values taken from local measurements (Table 1). Surface roughness below the forest canopy was taken as 0.2 m, reflecting a bracken understorey below the forest crown. The displacement height d of flow over the forest was calculated from (Massman and Weil 1999)

$$d/h=1-\int_0^1\left[-\overline{u'w'}\left(z\right)\Big/u_*^2\right]d\left(z/h\right), \tag{11}$$

where $-\overline{u'w'}\,(z)$ is the turbulent shear stress and u_* the friction velocity. Equation (11) is based on the assumption that d is the effective level of momentum drag on the canopy elements. Values of z_0 and d are sensitive to the forest structure parameter α, and a good fit to the observations was found for $\alpha = 5$ (Table 2).

Surface energy fluxes at every height in the forest have been assumed to be independent of fetch, since direct measurements (McAneney et al. 1994; Brunet et al. 1994; Todd et al. 2000) showed that, even under extreme advective conditions, the

Table 1 Input data, used for the simulation, averaged from K2002

Fluxes (W m^{-2})	$R_n - G$	H	LE
Bog	265	123	142
Forest	329	175	210
Other input data			
Forest height		19.7 m	
Measurement height at forest site		27 m	
Measurement height at bog site		8 m	
Average fetch over forest		256 m	
Roughness length forest (with long fetch)		2.1 m	
Displacement height (with long fetch)		12.5 m	
Leaf area index forest		1.8	
Roughness length bog		0.05 m	

Table 2 Roughness length (z_0) and zero-plane displacement (d) versus forest structure α

α	z_0 (m)	d(m)
1.5	2.6	7.7
3	2.4	10.3
5	2.1	12.4
9	1.7	14.6
13	1.4	15.7

surface flux showed hardly any variation with fetch, as explained by negative feedbacks between humidity deficit, stomatal conductance and transpiration (Itier et al. 1994; Baldocchi and Rao 1995).

3 Observations

For many atmospheric fluxes (e.g. CO_2, sensible heat, latent heat, momentum) it is hard to determine the corresponding surface flux. However, the sum of sensible and latent heat flux can be related to the energy absorbed at the surface using energy balance closure. Energy balance closure as a function of fetch downwind of a forest edge has been measured by Klaassen et al. (2002), referred to as K2002 in the following. Observations were made above a forest 150 m from a bog–forest transition on flat terrain. ‘Fetch’ is here defined as the distance between the measurement location and the forest edge into the wind direction, so variations in fetch arise from variations in wind direction. The data are presented as normalised energy flux (N), defined as:

$$N = \frac{H + LE}{R_n - G}. \tag{12}$$

Equation (12) shows the ratio between turbulent ($H + LE$, where H is sensible heat flux, LE is latent heat flux and L is latent heat of vapourisation of water) and non-turbulent fluxes ($R_n - G$, where R_n is net radiation and G is surface soil heat flux). In the case of equilibrium and perfect measurements of all energy fluxes, turbulent fluxes should equal non-turbulent fluxes, so $N = 1$ over forest. For fetches exceeding 400 m, K2002 found $N = 1.03 \pm 0.11$, or statistically not deviating from unity, but for fetches between 150 and 400 m, $N = 1.16 \pm 0.06$ is characteristic of the enhancement of turbulent fluxes. The enhancement of turbulent fluxes for short fetches has been explained by advection (K2002), yet surface heat fluxes for the upwind bog were even smaller than surface fluxes for the forest, implying that horizontal flux advection cannot explain the observed heat fluxes. Air temperatures above the forest and bog were almost equal but water vapour concentrations were generally higher over the bog. High water vapour concentrations over the bog occurred, not so much because of small surface evaporation but because of the low surface roughness (K2002). The small difference in behaviour of temperature and humidity, when the air moves from bog to forest, agrees with a slightly larger Bowen ratio over the bog (0.87) as compared to the forest (0.83). A decrease of water vapour concentration, when air moves from bog to forest, is consistent with an increase of latent heat flux with height, in accordance with the observed enhancement of turbulent heat fluxes.

4 Results

The normalised energy flux (N) is less than 1 for negative fetches as fluxes over the upwind bog are smaller than forest surface fluxes (Fig. 1). Relatively low energy fluxes over the bog resulted mainly from high ground heat flux (G) in the wet bog. Both measurements and simulations show $N > 1$ for fetches between 200 and 500 m downwind of the forest edge. The simulations show a gradual decrease of N towards unity for larger fetches, whereas the measurements suggest $N < 1$ for fetches around 700 m. A closer look at the measurement location revealed that, for fetches around 700 m, air flow was over an irregular forest patch with low tree heights. The complex forest structure in the upwind direction may cause the difference between observations and the simulation around 700 m fetch. Turbulent fluxes for moderate fetches (i.e. 200–500 m) are not only enhanced compared to the forest surface fluxes but also, and even more so, compared to the upwind surface fluxes. To analyse the influence of our upper boundary, a simulation was made with an extended vertical grid and double the boundary-layer height. The result, not shown in a figure, was less than a 5% change in energy fluxes at the measurement level for fetches exceeding 100 m, showing that our results are not sensitive to the choice of upper boundary.

All following simulations have been executed with upstream surface fluxes equal to forest fluxes. A constant surface flux over the full integration domain implies that surface forcing can be placed outside the integral of Eq. (1) and the integrated footprint (Eq. (2)) is then found. Constant surface fluxes result in a further enhancement of atmospheric fluxes over the forest (Fig. 2). Figures 1 and 2 show decreased fluxes for short fetches (<100 m or 5 times the forest height) and enhanced fluxes for larger fetches. Decreased fluxes imply negative vertical advection and enhanced fluxes imply positive vertical advection as explained in Sect. 2.1. The decrease of atmospheric scalar fluxes just behind the forest edge is caused by a decrease in mixing length due to a locally enhanced dissipation rate of TKE (Eq. (7)). In the following we will focus on the results for larger fetches where a closer relation between atmospheric fluxes and forest surface fluxes is to be expected. At moderate fetches, the simulations result in a reduction of wind speed inside and just above the forest, resulting in upward air movement and positive vertical advection.

The simulated footprint of atmospheric measurements above the forest at the measurement height $1.35h$ (27 m) is shown in Fig. 3 for two different locations downwind of the forest edge, at 400 m downwind of the edge and at infinite distance from

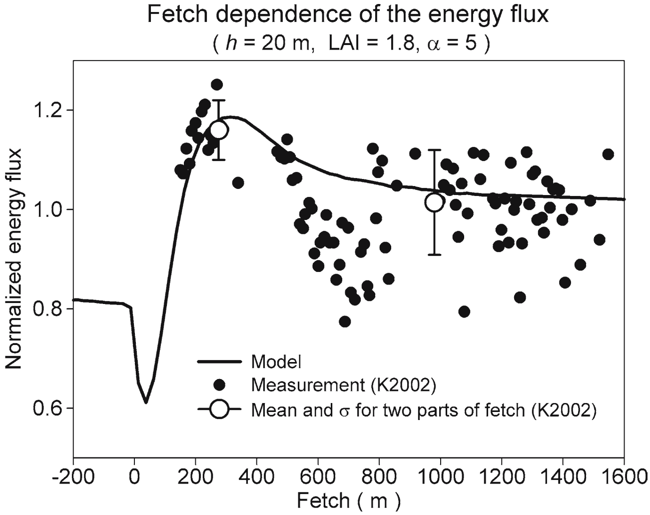

Fig. 1 Normalised energy flux at $1.35h$ (being the forest height, i.e. 27 m height) versus fetch downwind of the forest edge h Negative fetch values are upwind of the forest edge

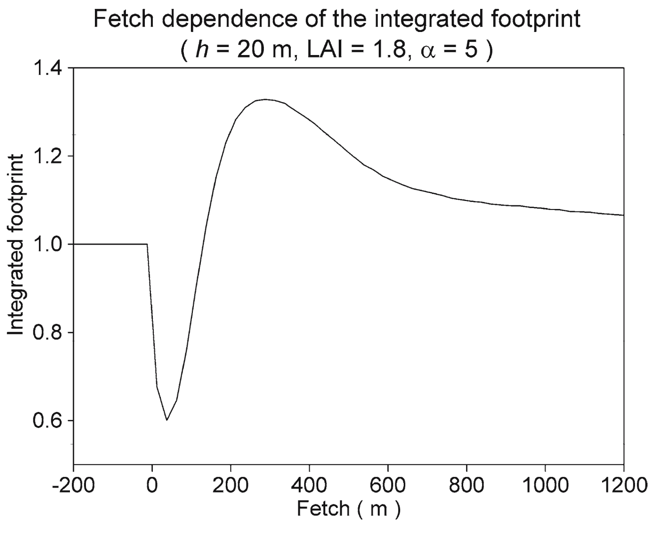

Fig. 2 Simulated integrated flux footprint at $1.35h$. This figure is derived using unit values of surface fluxes

the edge, representing homogeneous forest. The footprint over forest appears to be almost independent of measurement location, but the contribution of the upwind bog is increased as compared to the footprint of homogeneous forest. The small peak between 400 and 500 m upwind distance shows a peak in the contribution of the bog just upwind of the forest to the atmospheric flux over the forest at 400 m downwind of the edge. Enhancement of the upwind surface flux contribution to atmospheric fluxes over the forest is caused by enhanced mixing over the forest as compared to mixing over the aerodynamically smoother bog.

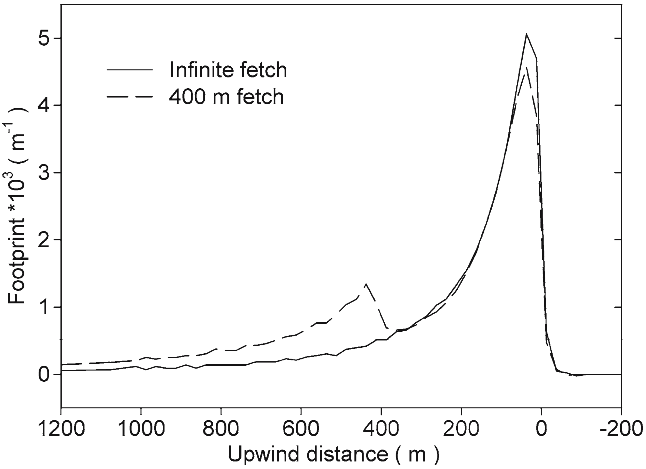

Fig. 3 Simulated flux footprint at $1.35h$. for two fetches downwind of the forest edge. 'Infinite fetch' shows the result for homogeneous forest and 400-m fetch, the result when measurements are taken 400 m downwind of a forest edge

Figure 4 shows the footprint integrated between infinite upwind fetch and the displayed upwind distance. The forest edge is found for the upwind distance equal to the fetch and the integrated footprint F_R is found when the upwind distance is zero. Figure 4 shows that, for homogeneous forest (infinite fetch), a unit integrated footprint is simulated, as expected, but for a measurement location at 400 m downwind of a forest edge the integrated footprint exceeds unity due to vertical advection. The secondary peak in Fig. 4 results from the relatively small energy fluxes at short fetches over the forest, see Fig. 2. The results of Fig. 4 are used to estimate the total contributions of bog and forest to an atmospheric measurement at 400 m fetch downwind of the forest edge: the integrated footprint over upwind bog $F_u \approx 0.5$ and over forest $F_i \approx 0.8$, implying for the atmospheric flux: $\eta(400\,\text{m}) \approx 0.5Q(\text{bog}) + 0.8Q(\text{forest})$. Conversely, the forest flux can be calculated from the integrated footprint using Eq. (4): $Q(\text{forest}) \approx 1.2\eta(400\text{m}) - 0.6Q(\text{bog})$. Note the strong deviation of the sum of the constants (1.2 – 0.6) from unity in this situation where the integrated footprint deviates from unity.

Figure 5 shows the normalised flux in the case of a constant surface source strength, which equals the integrated footprint (Eq. (1) with Q independent of r') in two dimensions. The figure shows fluxes slowly decreasing with height in the atmospheric surface layer over bog as is commonly observed during daytime. Atmospheric fluxes over forest show an area of enhancement with features of a plume escaping from the forest around 200 m fetch with a maximum around 300–400 m fetch (= 15–20 h) at the height of $1.5h$. It should be noted that assuming constant atmospheric fluxes with height over bog results in a slight further enhancement of fluxes over forest (see e.g. the difference between Figs. 1 and 2), and thus the process of flux enhancement over forest occurs independently of our particular choice of lateral boundary condition over bog.

The ratio (u_*/u) is taken as a measure for adjustment of the momentum flux, and therefore the values are normalised to the value for large fetch over forest (at the right boundary of the model domain). As wind speed u increases with height, we show only horizontal variations at the measurement height of $1.35h$. Normalised friction velocity

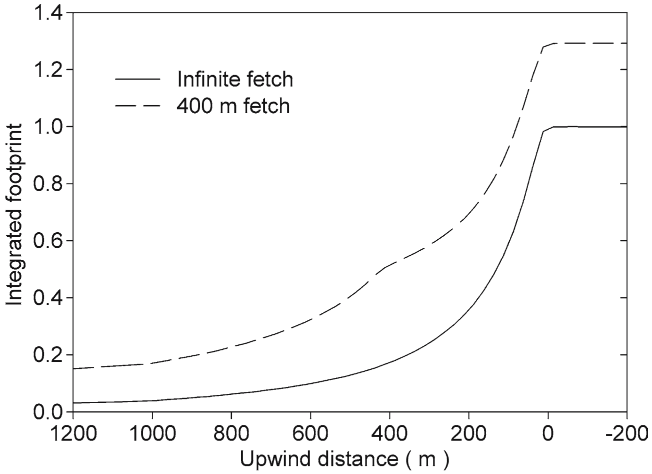

Fig. 4 Simulated integrated footprint at $1.35h$ for two fetches downwind of the forest edge

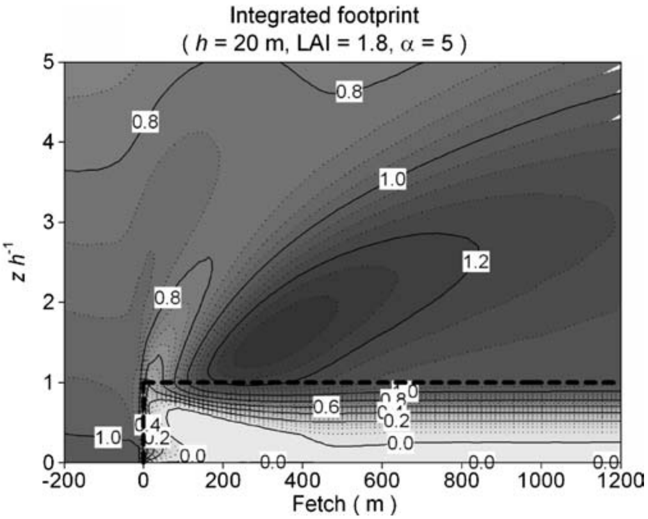

Fig. 5 Integrated footprint versus height and fetch

for $\alpha = 5$ adjusts within 10% at 300 m fetch (Fig. 6), in agreement with observations (Gash 1986; Kruijt 1994; Gardiner et al. 1995; Irvine et al. 1997; Van Breugel et al. 1999).

A sensitivity analysis is executed to test whether enhanced atmospheric scalar fluxes might be a common feature downwind of a forest edge. The general result of the sensitivity analysis is that height and fetch for enhanced fluxes are hardly sensitive to wind speed and scale with forest height (not shown in figures). Figure 7a shows that the maximum enhancement of turbulent flux increases with increasing forest density, but the fetch of enhancement decreases, as stronger coupling to the denser forest results in a concentration of the plume of enhanced turbulent fluxes to a smaller area downwind of the edge. Figure 7b shows that an increase of forest

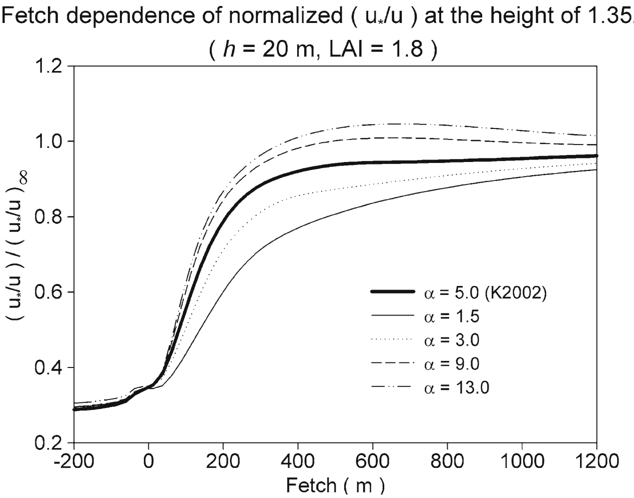

Fig. 6 Normalised friction velocity at $1.35h$ versus fetch as a function of forest structure

density increases the footprint over forest and decreases the value over the upwind bog, so atmospheric fluxes are more strongly coupled to denser forest. Forest roughness is found to decrease with increasing upper canopy density due to decreasing the exchange between atmospheric layers above and below the canopy crown. As a result, a dense upper canopy decreases the mixing of upwind properties to larger heights and decreases the contribution of the upwind area to the atmospheric fluxes for fetches less than 400 m (Figure 8b). The influence of leaf area density near the top of the forest canopy hardly affects the fraction of atmospheric flux arising from forest, as any increase of nearby leaf area is compensated by a decrease of atmospheric mixing. A significant (>10%) enhancement of integrated footprint is found for all forest structure parameters (Figs. 7a and 8a) around 400 m fetch.

5 Discussion and conclusions

Atmospheric flux measurements provide information on surface fluxes in an extensive upwind area, and it is commonly assumed that turbulent fluxes in the atmospheric surface layer equal average surface fluxes in the footprint area. Here we have shown that this assumption is not valid in the surface layer downwind of a forest edge, and atmospheric turbulent fluxes downwind of a forest edge are locally enhanced by perturbed turbulence and persistent upward vertical motion. The ratio between atmospheric fluxes and mean surface fluxes in the footprint area is given by the integrated footprint function.

The observed enhancement of turbulent heat fluxes over forest cannot be explained by horizontal advection alone, since heat fluxes from the upwind surface were even lower than heat fluxes from the forest. Thus, footprint models assuming homogeneous turbulence are not useful for estimating the forest flux from atmospheric measurements at moderate fetches (i.e. fetch–height ratios around 20) downwind of a forest edge. It is concluded that the turbulence field, in particular vertical advection, has to

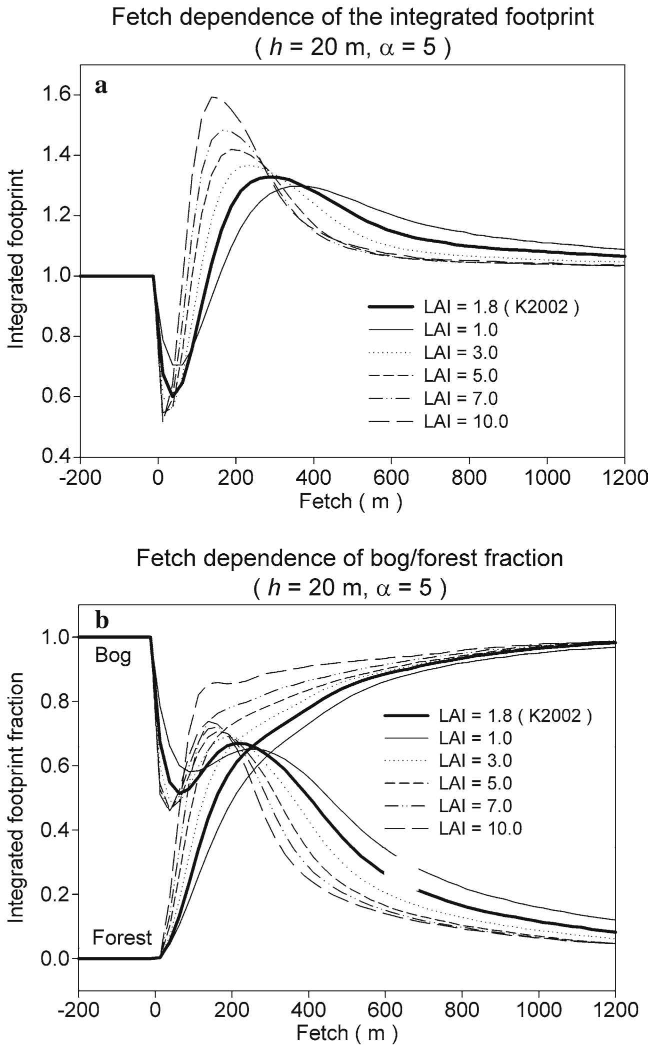

Fig. 7 Integrated footprint versus fetch at $1.35h$ as a function of forest density for: **(a)** the total turbulent flux, and **(b)** the flux fractions originating in forest and bog

be estimated as well in order to calculate forest fluxes from atmospheric flux measurements downwind of a forest edge.

Good agreement between simulated and observed atmospheric fluxes implies that forest fluxes can be calculated from corresponding atmospheric values using the footprint model presented here. It is concluded that this model is useful for calculating the footprint in partly forested landscapes with near-neutral atmospheric stability. This conclusion widens the conclusion of Sogachev et al. (2004) on the usefulness of the model in a landscape with moderate orography.

By comparing results for limited and infinite fetch, Fig. 3 shows that the footprint over forest is hardly dependent on measurement location, and suggests that common footprint models are useful for estimating the flux from the underlying surface

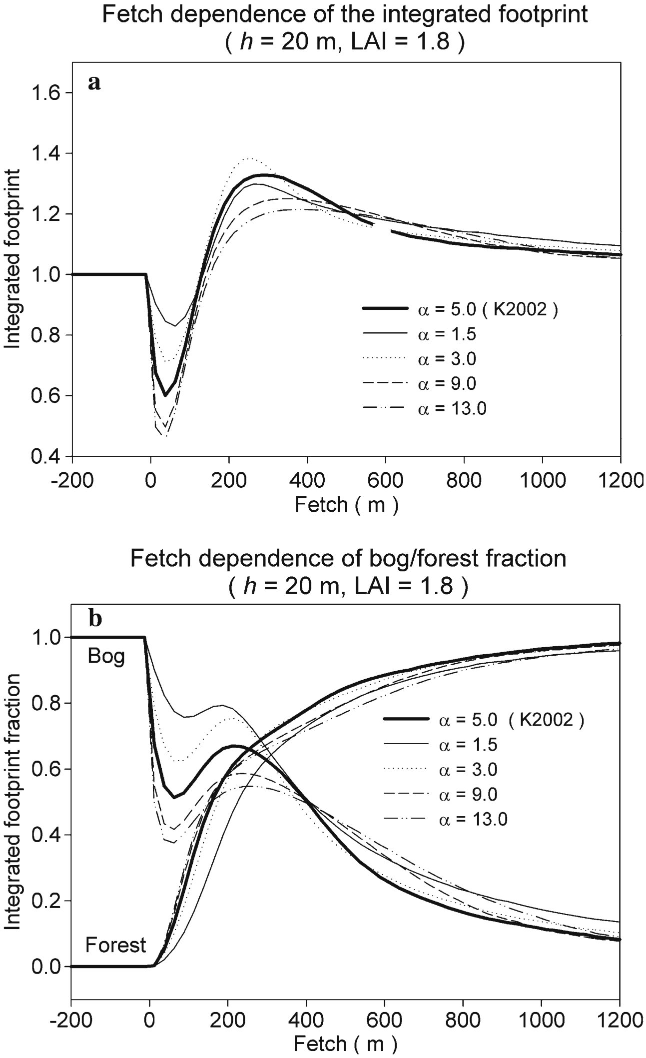

Fig. 8 Integrated footprint versus fetch at $1.35h$ as a function of forest structure for: **(a)** the total turbulent flux, and **(b)** the flux fractions originating in forest and bog

(horizontal advection). By contrast, the contribution of the upwind surface flux is strongly enhanced (a factor 2–3 in the analysed situation) as compared to a homogeneous situation. The enhancement is attributed to vertical advection, implying that the mean vertical wind $\overline{w}$, arising from a decrease of the horizontal wind by the aerodynamically rough forest, transports a quantity c (mass, momentum or heat) towards larger heights, see Eq. (6). It is recommended that field observations of atmospheric fluxes downwind of a roughness transition pay attention to vertical advection, and that footprint models take the actual turbulence field into consideration when they are used to analyse atmospheric flux measurements downwind of a forest edge.

The integrated footprint as a measure of atmospheric flux enhancement exceeds the value 1.1 for all forest densities and structures at moderate fetches over forest. It is concluded that the enhancement of turbulent fluxes is common downwind of a forest edge. Enhanced scalar fluxes extend to fetches up to 500 m, corresponding to a fetch: forest height ratio of 25, whereas normalised friction velocity tends to be adjusted within 300 m fetch (fetch:forest height ratio of 15). The more rapid adjustment of friction velocity is attributed to the low value over the upwind surface, so horizontal and vertical advection compensate to a large degree. It is concluded that the adjustment of friction velocity to the underlying forest does not imply an adjustment of scalar flux. It is recommended that more stringent fetch requirements for scalar flux observations be used, as compared to fetch requirements for momentum flux observations downwind of a roughness transition.

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References

Baldocchi DD, Rao KS (1995) Intra-field variability of scalar flux densities across a transition between a desert and an irrigated potato field. Boundary-Layer Meteorol 76:109–136

Brunet Y, Itier B, McAneney J, Lagouarde JP (1994) Downwind evolution of scalar fluxes and surface resistance under conditions of local advection. Part II: measurements over barley. Agric For Meteorol 71:227–245

Gardiner BA, Irvine MR, Hill MK, Baker M (1995) Airflow and turbulent flux development across a moorland/forest interface. Agric For Meteorol 21:71–73

Gash JH (1986) Observations of turbulence downwind of a forest-heath interface. Boundary-Layer Meteorol 36:227–237

Haenel HD, Grünhage L (1999) Footprint analysis: a closed analytical solution based on height-dependent profiles of wind speed and eddy viscosity. Boundary-Layer Meteorol 93:395–409

Horst TW, Weil JC (1994) How far is far enough?: the fetch requirements for micrometeorological measurement of surface fluxes. J Atmos Oceanic Tech 11:1018–1025

Irvine MR, Gardiner BA, Hill MK (1997) The evolution of turbulence across a forest edge. Boundary-Layer Meteorol 84:467–496

Itier B, Brunet Y, McAneney KJ, Lagouarde JP (1994) Downwind evolution of scalar fluxes and surface resistance under conditions of local advection. Part I: a reappraisal of boundary conditions. Agric For Meteorol 71:211–225

Klaassen W, Van Breugel PB, Moors EJ, Nieveen JP (2002) Increased heat fluxes near a forest edge. Theor Appl Climatol 72:231–243

Kormann R, Meixner FX (2001) An analytical footprint model for neutral stratification. Boundary-Layer Meteorol 99:205–226

Koster RD, Suarez MJ (1992) Modeling the land surface boundary in climate models as a composite of independent vegetation stands. J Geophys Res 97(D3):2697–2715

Kruijt B (1994) Turbulence over forest downwind of an edge. PhD thesis, University of Groningen, The Netherlands

Leonard A (1974) Energy cascade in large-eddy simulations of turbulent fluid flows. Adv Geophys A 18:237–248

Liu J, Chen JM, Black TA, Novak MD (1996) $E-\varepsilon$ modeling of turbulent flow downwind of a model forest edge. Boundary-Layer Meteorol 77:21–44

Markkanen T, Rannik Ü, Marcolla B, Cescatti A, Vesala T (2002) Footprints and fetches for fluxes over forest canopies with varying structure and density. Boundary-Layer Meteorol 106:437–459

Massman WJ, Weil JC (1999) An analytical one-dimensional second-order closure model of turbulent statistics and the Lagrangian time scale within and above plant canopies of arbitrary structure. Boundary-Layer Meteorol 91:81–107

McAneney KJ, Brunet Y, Itier B (1994) Downwind evolution of transpiration by two irrigated crops under conditions of local advection. J Hydrol 161:375–388

Pasquill F, Smith FB (1983) Atmospheric diffusion, 3rd ed. Wiley, New York, pp 437
Sanz C (2003) A note on k-ε modelling of vegetation canopy air-flows. Boundary-Layer Meteorol 108:191–197
Schmid HP (2002) Footprint modeling for vegetation atmosphere exchange studies: a review and perspective. Agric For Meteorol 113:159–183
Sogachev A, Lloyd J (2004) Using a one-and-a-half order closure model of the atmospheric boundary layer for surface flux footprint estimation. Boundary-Layer Meteorol 112:467–502
Sogachev A, Rannik Ü, Vesala T (2004) Flux footprints over complex terrain covered by heterogeneous forest. Agric For Meteorol 127:143–158
Sogachev A, Panferov O, Gravenhorst G, Vesala T (2005) Numerical analysis of flux footprints for different landscapes. Theor Appl Climatol 80(2–4):169–185
Todd RW, Evett SR, Howell TA (2000) The Bowen ratio-energy balance method for estimating latent heat flux of irrigated alfalfa evaluated in a semi-arid, advective environment. Agric For Meteorol 103:335–348
Van Breugel PB, Klaassen W, Moors EJ (1999) The spatial variability of turbulence above a forest. Theor Appl Climatol 62:43–50