Forest edges and the soil–vegetation–atmosphere interaction at the landscape scale: the state of affairs

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Abstract: Although the soil–vegetation–atmosphere exchange of momentum and heat is fairly well understood for many types of homogeneous surfaces, the disturbances created by transitions of one surface type to another remain to be analysed more fully. This is especially true for the impact which a large transition such as the forest edge has on the average fluxes in a small-scale heterogeneous landscape with forest. Recently acquired experimental evidence appears to some extent contradictory and at variance with conventional concepts.

Key words: forest edges, momentum fluxes, energy balance, land surface heterogeneity, advection, area averaging of fluxes.

1 Introduction

Observations on the water balance of forests have shown that the generally high water use of forests – as compared to, for example, low agricultural crops – is mainly caused by high evaporative losses of intercepted rainfall. One-dimensional modelling, reviewed by Veen and Dolman (1989), suggests that a high evaporation rate of intercepted water is explained to a large extent by the high aerodynamic roughness of forests. This can be easily understood. Strong absorption of the kinetic energy of the air flowing over the rough canopy creates strong turbulence. The upward motions in the eddies very quickly transport water vapour away from the wet, evaporating surface. When the leaves are dry, turbulence is no less intensive. But in a dry canopy the stomata

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regulate the transpiration rate. The surface resistance of forests is, therefore, much larger than the aerodynamic resistance. As in low vegetation it increases with decreasing availability of – among others – light and soil moisture. Water losses through transpiration in forests have turned out to be similar to those of other vegetation types.

The notion that forests present a very rough surface to the flow of air is important not only for the fluxes of water vapour but also for the other quantities that are transported by turbulent diffusion. The downward motions in eddies of the strongly turbulent wind field over forests transport gases and aerosols towards the land surface more efficiently than in the much quieter flow patterns across smooth, i.e., low vegetation. The flux of acidifying air pollutants from the atmosphere towards the forest soil is, therefore, significantly larger than the flux to arable soils. Measurements and calculations, for example, in the Dutch ACIFORN project (Van Aalst and Erisman, 1991) have demonstrated this.

Aerodynamic roughness also plays a role in the influence of forest on climate. The direct effect of forests on the atmosphere stems from high roughness in combination with low albedo. Low albedo causes much radiative energy to be available for conversion into sensible and latent heat. High roughness makes for low resistance to heat fluxes from the forest to the atmosphere. Extensive regions covered by forest significantly affect the properties of air masses overhead. At the continental scale the strong friction between atmospheric motion and extensive forested regions may even affect the general circulation patterns (Sud et al., 1988). The climatic importance of forest is further illustrated by various modelling studies of the climate impact of deforestation of Amazonia. Complete deforestation is expected to lead to a drier and hotter climate in the Amazon basin (Dickinson and Henderson-Sellers, 1988; Shukla et al., 1990; Lean et al., 1995).

In the aerodynamic sense, forest is the roughest type of vegetation surface available. However, intuitively it is obvious that the presence of forest edges contributes significantly to the drag forces which act upon the flow of air. Beljaars (1982), Wieringa (1992), Mahrt and Ek (1993) and Wang and Klaassen (1995) have provided experimental evidence that average, effective roughness of the land surface is indeed increased by the presence of tree rows, vegetation edges and other obstacles to wind flow. An increase of one order of magnitude is not uncommon. Taking the argument one step further, it seems possible that terrain with many small forests (<1–10 km²), and thus a high frequency of forest edges, may present a surface which is rougher still than the roughest vegetation component. On the other hand, in a terrain with large united blocks of forest (>100 km²) alternating with large, aerodynamically smooth vegetation, the frequency of edges may be too low to contribute noticeably to the area-averaged roughness. That is one reason for distinguishing vegetation heterogeneity at the ‘regional scale’, with a variety of homogeneous vegetation units of say 10 by 10 km or more, and the ‘landscape scale’, with a fine-grained pattern of smaller vegetation units (length scales of 0.1–10 km). Moreover, at the regional scale the planetary boundary layer (PBL) responds to the differences of the vegetated land surface in an organized manner. Owing to the different properties of the PBL over the different surfaces a mesoscale circulation may result, which will affect the area average of the heat fluxes. Thus, a special approach to flux averaging is indispensable (Shuttleworth, 1988). HAPEX-type experiments and meteorological mesoscale modelling are the tools with which this problem has been tackled (André et al., 1990). At the landscape scale no organized response of the PBL is noticed. Because of this it has been suggested that flux averaging at the landscape scale can be done by simply adding the contributions from the vegetation units.
(Shuttleworth, 1988). However, if the standard values of the roughness of the various vegetations are applied then such a strategy will yield inaccurate results. Especially when large roughness jumps such as forest edges are involved, the inaccuracy might be quite significant (Shuttleworth, 1994).

This article focuses on the landscape scale of the soil–vegetation–atmosphere exchange. After a brief sketch of flux-gradient relations over homogeneous vegetation (Section II) an outline will be given of the classical ideas on the flow of air across vegetation boundaries (Section III). Section IV reports on the rate at which the thickness of a new equilibrium internal boundary layer downwind of a roughness change builds up. This rate is found to be one order of magnitude larger over forest as compared to lower vegetation. This means that the transitional zones of modified flow downwind of forest edges are relatively very narrow. This could lead one to conclude that the number of edges, and consequently the scale of vegetation heterogeneity, does not greatly influence the average fluxes from heterogeneous landscape with forests. Such a conclusion is contrary to the argumentation presented above, as well as contrary to the theoretical study by Klaassen and Claassen (1995), and the observations reported by Kondo and Yamazama (1986), Chen and Schwerdtfeger (1989) and Dolman et al. (1994). From this apparent contradiction it follows that it is useful to have more insight in what exactly happens when air flow crosses a forest edge. As observed pointedly by Kaimal and Finnigan (1994), most existing two-dimensional theory on the behaviour of turbulent fluxes over changing vegetation goes back on a very few basic data sets, collected in the 1960s. It is, therefore, pertinent to review three new experimental studies on the effect of forest edges on the fluxes. This is done in Section V. The modelling approach is summarized in Section VI. Section VII analyses the meaning of the new experiments in relation to the standard ideas expressed in Section III. Some comments are given on the possible existence of a zone of maximum momentum flux downwind of a forest edge, as well as on the impact of the forest edge on heat and water budgets. It is concluded in Section VIII that we have advanced in our understanding of the mechanisms involved, but that the real impact of forest edges on the soil–vegetation–atmosphere interaction at the landscape scale can be finally evaluated only after more measurements have become available, which agree among them and with two-dimensional theory.

II Flux–gradient relations over homogeneous vegetation

As is known from observation, the deceleration of the wind near the land surface in the neutrally stratified atmosphere generally results in the logarithmic wind profile:

\[
\begin{align*}
\textit{u(z)} & = (u_*/k) \ln[(z-d)/z_0] \\
\end{align*}
\]

where

\[
\begin{align*}
\textit{u} & = \text{horizontal wind velocity} \\
\textit{u}_* & = \text{friction velocity} \\
\textit{k} & = \text{Von Kármán constant} \\
\textit{z} & = \text{height} \\
\textit{d} & = \text{zero plane displacement} \\
\textit{z}_0 & = \text{roughness length}.
\end{align*}
\]
Consider a small volume of air flowing with a certain horizontal velocity \( (u) \) at a level \((z)\) where a non-negligible change of wind speed with height occurs. Suppose that the turbulent air stream transports this volume downwards over a distance equal to the average diameter of the eddies occurring at that level. The volume of air will adjust to the lower horizontal wind speed of its new surroundings, i.e., its excess kinetic energy becomes available for further downwards transport towards the vegetation, where it is finally absorbed. This process results in the momentum flux from atmosphere to vegetated surface. In the PBL the momentum flux generally decreases with height. In the surface layer, the lower 10% of the PBL, the decrease with height of the momentum flux is assumed to be negligible. Therefore, over homogeneous vegetation the momentum flux \( \tau \) (otherwise known as shearing stress) in the atmospheric surface layer is considered to be constant with height. It is given by

\[
\tau = \rho K_M \frac{du}{dz}
\]

where \( \rho \) is air density, \( K_M \) is the constant of the proportionality of the (turbulent) momentum flux with the gradient, called (turbulent) diffusivity or eddy viscosity. Turbulent flow is often described using the friction velocity \( u_* \), which is given by \( \sqrt{\tau/\rho} \). Close to the earth’s surface, over homogeneous vegetation, the diffusivity \( K_M \) is in first approximation proportional with height: \( K_M = u_* k z \). The assumption of a constant momentum flux in the surface layer over an extensive, homogeneous surface is quite sound. Also, with increasing height in the surface layer the increase in diffusivity is largely compensated by the decrease of \( du/dz \), even with small variations in \( \tau(z) \). With current technology the momentum flux can be measured directly by recording the fluctuations of wind speed in the \( x, y \) and \( z \) direction, using fast-response sonic anemometers. This direct observation of turbulence offers advantages over the earlier method of deriving roughness parameters from vertical wind speed profiles, obtained with standard anemometers (Lloyd et al., 1992). The latter method is sensitive to errors (Wieringa, 1993).

The fluxes of quantities such as water vapour, other gases and sensible heat are linked to the turbulent transport of momentum. In the surface layer over homogeneous vegetation these fluxes are also expected to be constant. At the bottom of this constant-flux layer, turbulent flow is affected by the geometry of the vegetation: the roughness sublayer. In the constant-flux layer above the roughness sublayer the best conditions prevail for measuring fluxes representative of the underlying surface.

The equations connecting the fluxes of heat and gasses to the gradients involved are

\[
H = K_H \left( \frac{dT}{dz} \right)
\]

\[
E = K_V \left( \frac{dq}{dz} \right)
\]

\[
F = K_S \left( \frac{dc_s}{dz} \right)
\]

where \( H \) and \( E \) are sensible and latent heat flux respectively, \( T \) temperature, \( q \) humidity deficit of the air, \( c_s \) concentration of the trace gas, and \( K_H, K_V \) and \( K_S \) the diffusivities for sensible heat, water vapour and the trace gas under consideration.

Equations (2) to (5) describe the atmospheric fluxes towards and from extensive and homogeneous land surfaces under conditions of neutral stability in the lower atmos-
The simplest approach to the relationship between the fluxes is to assume that the diffusivities are equal:

\[ K_M = K_H = K_V = K_S \]  \hspace{1cm} (6)

Direct observations of the fluxes, using fast sensors and the eddy correlation approach, have demonstrated that in fact the diffusivities may not be equal. It is useful to correct for their differences (Lang et al., 1983). With non-neutral stability of the atmosphere, Equations (2) to (5) remain valid, but diffusivity is no longer equal to \( u_* k z \). Reasonably satisfactory stability functions can be added to the flux equations (see, e.g., Brutsaert, 1982).

### III Air flow across vegetation boundaries

Consider the case where air has travelled over a sufficiently long stretch of homogeneous land surface to have acquired profiles of wind speed, temperature and humidity which are in equilibrium with the fluxes of momentum, sensible heat and water vapour from the underlying surface. When the flow crosses a vegetation boundary, the air adjusts to the roughness and wetness of the new underlying surface. After a certain distance beyond the transition new profiles of wind speed, temperature and humidity will be established, reflecting the fluxes from the new surface. Figure 1 shows the general case for a smooth-to-rough transition, as it is conventionally presented.

![Figure 1](image_url)

**Figure 1** (a) Development of a new internal boundary layer when air moves from a smooth vegetation (S) towards and over a rough vegetation (R) (after Monteith and Unsworth, 1990). Note that the momentum flux in the zone of modified flow (\( \tau' \)) is shown to be intermediate between those of the smooth and the rough surface (\( \tau_S \) and \( \tau_R \) respectively); (b) behaviour of mean wind speed \( \bar{u} \) and surface shearing stress (\( \tau'_0 \), equivalent to surface momentum flux) when air flow passes a smooth–rough transition (after Oke, 1978). Note the overshoot of \( \tau'_0 \) just downwind of the edge.
The momentum flux over the relatively smooth surface, $\tau_S$, is smaller than the momentum flux over the relatively rough surface, $\tau_R$, since the latter presents a rougher surface. The figure suggests that the momentum flux in the transition zone of modified flow is intermediate between $\tau_S$ and $\tau_R$ (Figure 1a). As mentioned in Section I, the presence of a roughness jump causes additional drag forces. This effect is shown in Figure 1b, where the surface shearing stress $\tau_{OR}$ exhibits a maximum (the ‘overshoot’) just downwind of the edge. This means that the surface just downwind of the edge acts as an extra source of momentum. Usually, local (point) sources produce a downwind plume of the emitted quantity (Pasquill, 1974). Therefore, one would expect that this overshoot causes at least locally a zone in which the momentum transfer in the zone of modified flow surpasses $\tau_R$. In Figure 1a there is no indication of this.

When air is advected from one type of land surface to another, temperature will generally also be affected. For a ‘dry’ to ‘wet’ change in surface conditions, it is usually assumed – on the strength of the measurements by Rider et al. (1963) – that dry advected air will cause the evaporation downwind of the change to overshoot. A shift then occurs in the ratio of latent to sensible heat transport: the lowest air layers cool down, and a downward flux of sensible heat from higher air layers results. In the field conditions of the measurements by Rider et al. (1963) the energy balance

$$R_n - G = H + \lambda E$$

(with $R_n$, net radiative energy, $G$ ground heat flux, $H$ sensible heat and $\lambda E$ latent heat) still closes downwind of dry–wet transition. Recently, Itier et al. (1994) have argued that often the tendency for evaporation to overshoot at an edge which is subject to advection of relatively warm and dry air may be counteracted by a feedback of stomatal behaviour. When a dry to wet change is accompanied by a large roughness change a different effect occurs. Upwind of the change, the temperature and humidity profiles are adjusted to the smooth surface and produce large gradients, but relatively high atmospheric resistance precludes large fluxes of heat and humidity. Downwind of the transition increased turbulence causes lower resistance, which leads locally to higher fluxes. This way $(H + \lambda E)$ may surpass $(R_n - G)$. In the area affected by heat advection, the one-dimensional approach of considering flux–gradient relations only in the vertical direction breaks down, since that approach is based on the validity of the energy balance (Equation 7).

IV The problem of fetch

A flux sensor mounted above a certain vegetation type has a ‘field of view’ which is a function of sensor height. The higher the sensor is mounted, the larger the contribution will be from sources further upwind. If the objective is to measure the flux characteristic of the underlying vegetation, it is necessary to have specific information on the relation between sensor height and the upwind distance of the source area to the sensor. Traditionally, rules of thumb have been applied for estimating the fetch required to obtain values of the fluxes representative of the underlying surface, so as to justify the application of one-dimensional theory. According to Oke (1978) the fully adjusted internal boundary layer grows rather slowly, requiring a fetch distance of 100–300 m for every one metre increase in the vertical. Such figures are based on the flow
equations predicting the buildup of a disturbed layer downwind of the roughness jump in which the lower 10% is expected to be in equilibrium with the new surface. A number of micrometeorological studies have confirmed their validity for surfaces with either grass or low crops (e.g., Munro and Oke, 1975; Arya and Shipman, 1981). However, a correction is needed when taller, i.e., rougher vegetation is involved. A simple approach which includes the important effect of surface roughness has been put forward by Gash (1986b):

$$X_F = \frac{-z \left( \ln \left( \frac{z}{z_0} \right) - 1 + \frac{z_0}{z} \right)}{k^2 \cdot \ln(F/100)}$$

(8)

$F$ is the percentage of the flux originating from the source at the distance $X_F$ from the sensor, while $z$ is the height of measurement above the zero plane displacement height $d$. A uniform wind field has been assumed. Applying Equation (8) to a water vapour flux measurement at $z = 2$ m over short grass with $z_0 = 0.01$ m one finds that 10% of the measured flux originates between the sensor location and a distance of 22 m, 50% from between 0 and 74 m and 90% from 0 to 486 m. The usual criterion for an adequate fetch is that 90% of the adjustment of profiles and fluxes to change of land surface has been completed. So when measuring fluxes at 2 m height over grass, a fetch of about 500 m yields data representative of the surface. This result is in agreement with the 1/100 to 1/300 rule.

Application of Equation (8) to forest yields a much shorter critical fetch. The displacement height $d$ in forest may be set to 0.75 of vegetation height $h$ (Rutter et al., 1971). Assuming a roughness length of 1 m, and a measurement height $z$ of 5 m above $d$ yields a 90% effective fetch of 230 m. Considering the ratio between height of the sensor and the 90% effective fetch, one calculates for a 10 m high forest at height/fetch ratio of 12.5/230 = 1/18 for the forest. Comparing this to the value calculated for grass (1/243), one notes that applying the rule-of-thumb estimate would overestimate the required fetch for forest considerably. Experimental support for large height/fetch ratios over forest was also obtained by Gash (1986a). Measuring windspeed components and friction velocities along a heather–forest interface (see also Section V), he concluded that traditional height/fetch ratios are much too conservative. His measurements indicate ratios of 1/27 to 1/16, in agreement with the calculations according to Equation (8). The height/fetch ratios for forest suggest that air flowing over a forest adjusts far more quickly than when flowing over a much smoother surface. As stated in Section I, this result seems to be an apparent contradiction to the information which points to a considerable effect of forest edges on average roughness, and thus on the magnitude of the fluxes. This motivates a closer look at what happens exactly near a forest edge.

V Fluxes observed downwind of a forest edge

Experimental studies of fluxes near a forest edge are rare. The first, pioneering, study in England by Gash (1986a) was mentioned in the previous section. His measurements relate to the flow across an interface between heather (predominantly Calluna) and a 9–11 m high forest with pines (P. sylvestris and P. nigra) and larch (Larix decidua). The
topography of the site is flat. The measurements were executed at a measuring height of 13.5 m above ground, so at 1.35 canopy heights (see also Table 1). Having performed a careful error analysis, Gash (1986a) inferred from his experimental data that the turbulence parameters did not significantly change between distances of about 12–40 canopy heights (120–400 m) downwind of the forest edge. The standard deviations in the measurements of this experiment are such that few inferences can be drawn as to the validity of the general picture of the momentum flux as shown in Figure 1a. The decrease of mean horizontal wind speed over the forest ($\bar{u}_R$ compared to $\bar{u}_S$ in Figure 1b) is confirmed.

More recently, Gardiner et al. (1994) reported on a forest edge experiment in Scotland, in which the edge separates pasture from a dense, 6–7.5 m high Sitka spruce plantation forest. The terrain slopes gently (~2°). The edge faces the prevailing wind direction, i.e. southwest. They measured momentum fluxes at distances from the edge of about 0.5, 5 and 15 times the canopy height respectively. More information on their experimental setup is summarized in Table 1. The largest momentum fluxes were recorded at the longest fetch, at canopy level. At that same fetch of 15 canopy heights the fluxes

Table 1  Comparison of experimental design of three studies on the edge effect of forests

<table>
<thead>
<tr>
<th></th>
<th>Height of forest (m)</th>
<th>Assumed average height of zero plane</th>
<th>Height of zero plane</th>
<th>Measurement height relative to ground level</th>
<th>Tree height of zero plane</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gash</td>
<td>9.2–11</td>
<td>10.0</td>
<td>7.5</td>
<td>13.5</td>
<td>6.0</td>
</tr>
<tr>
<td>Gardiner et al.</td>
<td>6–7.5</td>
<td>7.5</td>
<td>5.6</td>
<td>4.0</td>
<td>-1.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>7.5</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>15.0</td>
<td>9.4</td>
</tr>
<tr>
<td>SHEAR</td>
<td>22–23</td>
<td>22.0</td>
<td>15.0</td>
<td>26.0</td>
<td>9.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>33.0</td>
<td>16.5</td>
</tr>
</tbody>
</table>

b  Distances of measurements from the forest edge

<table>
<thead>
<tr>
<th></th>
<th>Distance in m</th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Gash</td>
<td>50</td>
<td>120</td>
<td>200</td>
<td>400</td>
<td>700</td>
</tr>
<tr>
<td>Gardiner et al.</td>
<td>4</td>
<td>30</td>
<td>112</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SHEAR</td>
<td>88</td>
<td></td>
<td></td>
<td></td>
<td>220</td>
</tr>
</tbody>
</table>

In relative fetch units ($x/h_{av}$)

<table>
<thead>
<tr>
<th></th>
<th>Distance in $x/h_{av}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gash</td>
<td>5</td>
</tr>
<tr>
<td>Gardiner et al.</td>
<td>0.5</td>
</tr>
<tr>
<td>SHEAR</td>
<td>4</td>
</tr>
</tbody>
</table>
were considerably lower at two canopy heights than at one canopy height, but they were still larger than the fluxes measured at the smaller fetches. This indicates that the representation of Figure 1b may be too simple. Rather than be constant both $\tau_R$ and $\tau_{OR}$ behave as a function of fetch. At all but the shortest fetch they found a clear decrease of the momentum flux with height. This result tends to confirm the distribution of momentum fluxes as schematized in Figure 1a. The value of the momentum flux as measured by Gardiner et al. (1994) at canopy level ought to be related to the surface shearing stress $\tau_{OR}$ as pictured in Figure 1b. However, in their canopy-level observations, the momentum flux behaves quite differently. Instead of an overshoot just inside the edge and a leveling off to a constant value, they observed a steady increase with fetch of the canopy momentum flux. So their observations do not support the idea that the edge itself contributes to extra momentum flux.

Some empirical evidence that the edge contributes to extra drag, and therefore extra momentum transfer, was found during a third experiment which ran in 1990 and 1991 in The Netherlands. Some design parameters of this Sleen Hydrometeorological Experiment on Advection and Regionalization (SHEAR) are given in Table 1. The SHEAR sites is on flat terrain. Land use is partly plantation forestry on sandy soils, partly agricultural.

Westerly winds prevail. Flux measurements were performed at several heights above a mixed deciduous forest, containing oak ($Q. \text{rubra}$ and $Q. \text{robur}$), beech ($Fagus \text{ sylvatica}$) and larch ($Larix \text{ leptolepis}$). Canopy height was 22 m. Fetch varied with wind direction from about 4 to 10 canopy heights. As will be seen in Figure 2, the length scale of the

Figure 2 Map of the SHEAR site. The broken lines at the left-hand side approximately delineate sectors of upwind surface type. Southwesterly to northwesterly winds arrive at the observation tower after crossing the well defined edge between the high forest from low crops, mainly grass (sector Gr). Northwesterly to northeasterly winds arrive after crossing a secondary edge between low and high forest (sector Lf); the upwind vegetation of the remaining sector is irregular in height and composition (sector Bad)
entire forest is in the order of 2 km, while individual parcels within the plantation typically are 0.2 km in diameter. Stands differ in composition, age and height. To the west of the forest the agricultural areas are predominantly grassland. To the east, crops are grown. In addition to the forest observations, fluxes were measured above grass and low crops (barley). Early results of the experiment were given by Hutjes et al. (1991), Kruijt et al. (1991) and Veen et al. (1991). Kruijt (1994) and Kruijt et al. (1995) have reported on details of instrumentation, data handling, error analysis and on observed vertical flux differences over the forest. Hutjes (1996) reports on the observations at two agricultural sites, and on the microclimatology of the various vegetation types.

The most useful SHEAR observations on momentum fluxes appear to be those measured at 4 and 11 m above the 22 m high canopy. Depending on wind direction, the fetch, \( x \), varied considerably. The relative fetches \( x/h \), where \( h \) is canopy height) ranged from about four to ten. In Figure 3, from Kruijt (1994), the ratio of the momentum transport at 11 m above the canopy to that at 4 m above the canopy is plotted.

**Figure 3** Observed difference of the momentum fluxes at \( z = 33 \) m (1.5 canopy height) and \( z = 26 \) m (1.2 canopy height), normalized on the \( z = 26 \) m flux, over forest, against relative fetch (in canopy heights). The diagram represents all SHEAR measurements from sector Gr (cf. Figure 2), under the three classes of atmospheric stability. Positive values indicate increase of flux between lower and upper level. Such an increase is observed upward from a distance from the edge of six to seven tree heights (after Kruijt, 1994)
against relative fetch (for the three atmospheric stability classes). Closer to the forest edge the momentum flux systematically decreased with height, while at the larger fetches ($x/h > 6$) the transport of momentum increased with height above the canopy. This not only means that the momentum flux in the surface layer in this situation is not constant with height but also that an increase of the flux with height is apparent at some distance of the edge. The latter observation does not fit into the general picture of the momentum flux in the modified zone being intermediate between the fluxes over the smooth and the rough surface (Figure 1a). The observation, however, is consistent with the idea that a plume of maximum momentum flux is generated downwind of the forest edge owing to the extra drag produced by the edge itself.

Apparent from the SHEAR measurements, furthermore, is a significant, local edge effect on the sensible and latent heat fluxes. Figure 4 shows a plot of the ratio of outgoing sensible plus latent heat over incoming net solar radiation ($\frac{(H + \lambda E)}{Q_n}$), for all wind directions (Hutjes, 1996). If the wind arrives from the grassland west of the forest edge, the plot indicates that the heat fluxes (sensible plus latent heat) given off by the forest may be up to 25% larger than incoming solar energy. This is far more than can be accounted for by the soil heat flux, which was measured during part of the experiment (Hutjes, 1996). If the wind has a longer fetch over the forest canopy, then the ratio $\frac{(H + \lambda E)}{Q_n}$ is mostly close to unity. Exceptions correlate roughly with secondary edges between low, young forest upwind and the high mature forest, where the tower was located.

A different picture emerges when the heat budget data from all wind directions are lumped. Figure 5 shows, despite the large scatter of individual observations, that $\frac{(H + \lambda E)}{Q_n}$ does not significantly deviate from unity. The implication of this is discussed in Section VII.

![Figure 4](ppg.sagepub.com) 

**Figure 4** Observations on the energy balance (sensible plus latent heat flux, divided by net radiation) over the SHEAR forest against wind direction. Note that in case the wind arrives from the smooth surface to the west (sector Gr) the excess of the heat flux over net radiation is as large as about 25% (after Hutjes, 1996)
Figure 5 Energy balance over the SHEAR forest, as observed over a prolonged period in which wind arrived at the forest tower from all directions. Apparently, local heat advection effects cancel out when energy balance data are lumped (after Hutjes, 1996).

VI Models of two-dimensional flow over heterogeneous vegetation

Many efforts have been made to model the two-dimensional pattern of momentum flux when air flows over roughness changes. The problem is that the two-dimensional flow equation cannot be solved analytically. The set of flow equations contains more unknown quantities than equations. One approach is to assume an expression for momentum diffusivity $K_M$ (Equation 1). The simplest way (first-order closure) is to equate $K_M$ to the product of the friction velocity $u_*$ and a length scale related to the eddy size, the mixing length, $l_m$, given by the product $kz$. Alternatively, the mixing length is expressed by functions relating it to the turbulent kinetic energy (TKE) fluctuations in the wind field (one-and-a-half order closure; Peterson, 1969; Taylor, 1970; Shir, 1972; Lo, 1986; Beljaars et al., 1987). Claussen (1988) has reviewed the impact of the TKE class of closure techniques on the properties of the modelled flow. A more complex approach (second-order closure) has been applied to the roughness change problem by Rao et al. (1974a; 1974b).

Generally, in these models the surface is considered as a flat surface, or ‘big leaf’ (Monteith, 1965; Thom, 1975). In such a representation, all surface drag is situated in this single layer, and is determined by the roughness length $z_0$. A roughness transition is then characterized by a sudden change in $z_0$. This approach neglects the enhanced exposure of the vegetation at the edge to the wind blowing on to the edge. Also, these big leaf models do not account for a change in zero-plane displacement, while such a change is considerable at forest edges. Li, Miller et al. (1985) and Li, Lin et al. (1990) have constructed a more realistic model, which allows not only for flow over but also into the forest edge. They included a multilayer representation of the forest biomass, pressure effects and first-order closure. In a further development of this approach Klaassen (1992) designed a multilayer model to estimate average fluxes over hetero-
geneous vegetation, characterized by smooth–rough transitions. He concentrated on modelling explicitly the variation of mixing lengths $l_m$ with downwind distance from the edge. Bearing in mind that mixing lengths are related to the sizes of turbulent eddies, he assumed $l_m$ to advect along the flow, and to adjust only slowly to a new surface. Adjustment rates for $l_m$ were determined by empirical parameters, calibrated against measurements of Bradley (1968), and validated on those of Gash (1986a).

Figure 6 shows the two-dimensional distribution of momentum fluxes downwind of a forest edge as simulated by the Klaassen (1992) model. At fetches larger than about seven canopy heights the model simulates a decrease of the flux from the canopy upward. At smaller fetches closer to the edge an increase with height is simulated up to about two canopy heights. Upward of this level the momentum flux is seen to decrease. This model result suggests the presence of a plume of maximum momentum flux, generated by the extra drag at the forest edge.

VII Discussion

From Section IV it follows that a flux sensor above vegetation downwind of a smooth-to-rough transition may 'see' part of the upstream smooth area, as well as an edge effect, and part of the rough area. In the measured flux the contributions from these three components are integrated. Local flux maxima are, therefore, not easily detected. Not even the analysis of the spectral distribution of the instantaneous fluctuations can solve this problem. The spatial integration inherent in the signal may be called the 'blurring effect', and it is basically the result of the nature itself of turbulent atmospheric transport. Three-dimensional source weight functions have been proposed for the relationship between the spatial distribution of surface sources and a signal measured in the surface layer. The integral of the source weight function over a specified domain is the 'effective fetch', or source area (Schmid and Oke, 1990; Schmid, 1994).

![Figure 6](https://example.com/figure6.png)

**Figure 6** Two-dimensional distribution of the momentum flux (in $m^2/s^2$) downwind of a 20 m high forest edge, as simulated by the Klaassen (1992) model. The heavy line outlines forest. Note the plume-like structure immediately downwind of the edge (after Kruijt, 1994)
As was shown in Table 1, the three studies on forest edges differ significantly in design. As a result, measurements were taken at quite different positions in the wind field. This is true for both absolute heights and distances as well as for heights and distances scaled with canopy height. This means that in each of the measurements of these studies the 'blurring effect' manifests itself in a different and hard-to-quantify manner. This makes a direct intercomparison of the results quite difficult. We therefore choose to compare the results with the general concepts as shown in Figure 1. The proposition that in the zone of modified flow the momentum flux is intermediate between the fluxes on both sides of a forest edge is neither supported nor undermined by Gash (1986a), and is consistent with the results published by Gardiner et al. (1994). If the momentum flux measurements at canopy level by Gardiner et al. are representative of the surface momentum flux (shearing stress) \( \tau_{OR} \), then the observed increase with fetch contradicts the behaviour of \( \tau_{OR} \) as shown in Figure 1, where it is constant downwind of the overshoot. From Gardiner et al.'s data it looks as if the surface shear is a function of fetch. The SHEAR experiment is the only one which yields an indication for an increase of momentum flux from the forest canopy upwards at some distance from the forest edge. Such an increase suggests that at least in some part of the zone of modified flow the momentum flux \( \tau \) is not intermediate between the values for the smooth and the rough surfaces, but in fact higher than the flux from the rough surface. It is tempting to interpret this as an indication for the existence of a plume of maximum momentum flux in the wake of the forest edge (Figure 7). Such a plume is impossible if there is no overshoot of \( \tau_{OR} \), the surface momentum flux. Since Gardiner et al. (1994) found no indication for an overshoot of \( \tau_{OR} \) (Section V), the Gardiner et al. and SHEAR results appear contradictory.

In a wind tunnel experiment, Kawatani and Sadeh (1971) observed a plume of maximum momentum flux downwind of the edge of a model forest. In a wind tunnel the upper boundary condition is a plane of zero diffusivity. In the real world such a condition is not present. Instead, above the upper boundary of the zone of modified flow a condition exists of diffusivity which is suppressed as compared to the condition in lower layers of air. This difference between the wind tunnel situation and the field

![Figure 7 Schematic representation of hypothetical plume of maximum momentum flux downwind of a forest edge (after Kruijt, 1994)](ppg.sagepub.com)
situation is the reason why the observation by Kawatani and Sadeh (1971) has to be considered support for, rather than proof of the occurrence of a momentum flux plume occurring in nature.

The big leaf models mentioned in Section VI do reproduce the overshoot of \( \tau_{OR} \) (Figure 1b), but not a plume of maximum momentum flux. The multilayer model by Klaassen (1992), allowing for advection of mixing length, is a notable exception. Still, its results do not fit the SHEAR data exactly. At around five relative fetch units the observations suggest a maximum momentum flux, whereas the model locates the maximum closer to the edge. Gardiner et al. (1994) did not measure at heights between one and two tree heights. Therefore, comparison with the model’s results remains inconclusive. Since Gash (1986a) measured at a single, fixed height over the canopy, his study does not yield information of flux divergence in the vertical at a fixed distance of the edge. This makes comparison with the model’s results not meaningful.

The SHEAR observations of the energy balance at the upwind forest edge confirm that local advection of heat may significantly increase the transfer of sensible and latent heat from the peripheral zone of the forest near the upwind edge towards the lower atmosphere. This result is consistent with the occurrence of a plume in which increased momentum transfer leads to high transport of sensible and latent heat. High transfer of latent heat over forest downwind of an edge should lead to an increase of interception loss. However, such an increase is not observed. This might be explained by lower canopy storage capacity of the forest near the edge (Klaassen et al., 1996). Not only is heat transported more quickly in the forest edge zone but also that zone receives significantly higher amounts of nutrient and acidifying deposition (Hasselrot and Grennfelt, 1987; Beier et al., 1992; Draaijers et al., 1994). The latter observation is usually interpreted in connection to the horizontal inflow of polluted air into the forest. It stands to reason that a possible turbulence maximum downwind of the edge would increase the downward fluxes of air pollutants in that area. The notion that the area just inside the forest edge is a special, high-flux environment agrees generally with the observations by ecologists and foresters, such as discussed by Chen et al. (1993) in their study of the microclimate in the forest edge itself.

While the energy budget at the forest edge is clearly unbalanced when winds arrive from a smooth upwind surface, the evidence produced by SHEAR indicates that this phenomenon is not noticeable when energy balance data are lumped over all wind directions. This must mean that heat advection at the leading edge and at the lee side of the forest more or less cancels out. This agrees with suggestions that heat fluxes are affected to a lesser extent by increased momentum fluxes because of the feedbacks on available radiative energy and water (Beljaars and Holtslag, 1991; Claussen and Klaassen, 1992; Hignett, 1994). Raupach and Finnigan (1995) recently suggested that a feedback on the Penman–Monteith evaporation at the surface could also be involved.

VIII Summary and conclusions

Whereas the momentum flux in the atmospheric surface layer over extensive homogeneous vegetations is constant with height, both experimental and model results indicate that this is not true for the situation downwind of a smooth-to-rough transition. It takes some time and distance for air flow to adjust to the rougher surface. The new, more turbulent internal boundary layer over the rougher surface builds up slowly. But
if the rough surface is forest, this buildup is not so slow compared to a lower vegetation. Still, over quite a distance downwind of the forest edge, the momentum flux closely above the forest canopy is much larger than over the forest at the larger height of a few tree heights. This decrease in the surface layer of the momentum flux with increasing height poses two problems. The magnitude of an observed flux is not only determined by wind speed and surface roughness but also by the measurement height. Secondly, one-dimensional theory is inappropriate for calculating the contribution of forest edges to the area-averaged momentum flux over forested landscapes. Two-dimensional theory predicts overshoot of the surface shearing stress just downwind of a forest edge, but atmospheric momentum fluxes in the transition zone that are intermediate between those of the smooth and rough surfaces. Comparing three recent micrometeorological forest edge experiments with theory reveals that the experimental results to some extent disagree with theory. In the fully adjusted layer the momentum flux at canopy level (which should be very close to the surface shearing stress $\tau_{OR}$) is not observed to overshoot at the leading edge. Neither is it constant with fetch, but it is observed to increase. If there is no overshoot in $\tau_{OR}$ there cannot be a plume of maximum momentum flux downwind, while some evidence for the existence of such a plume was found. If further substantiated, the existence of a plume disproves the validity of the notion that the momentum flux in the zone of modified flow is in all points intermediate between the fluxes over the smooth and the rough surface.

On the strength of these findings, the basic concepts of theory (Figure 1) cannot be considered to be fully disproved nor fully confirmed. This is caused by a) the limitations set by the designs of the experiments; b) the indirect way in which the surface shearing stress was measured; and c) the likely influence of the blurring effect in the measurements. So, the study of forest edges continues to deserve attention. There are additional reasons for this. Direct and circumstantial evidence points to the notion that forest edges are a special, high-flux environment. This has environmental and ecological implications. Last but not least, the large average roughness of heterogeneous terrain, with a high frequency of forest edges, must have an impact at the landscape scale on the momentum fluxes. Because they are transported by the same mechanism that transports momentum, i.e., the turbulent eddies, the area-average fluxes of other scalars must be affected. For the heat fluxes this impact is probably minor. The fluxes of gases are likely to follow the behaviour of the momentum flux more directly, and be enhanced by heterogeneity. For the fluxes of aerosols such enhancement has already been demonstrated in a number of studies.

The available studies on the effect of the forest edge on the soil–vegetation–atmosphere interaction have helped in identifying the relevant issue. A forest edge is not simply a roughness jump but it is also the upwind boundary of a zone in which the one-dimensional big leaf model of vegetation breaks down. As yet there is no good agreement between the two-dimensional theory and observations in this zone. Further experimental studies are needed to decide on the real importance of the forest edge effect on the fluxes at the landscape scale.

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