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Evaporative Conditions Across a Grass-Forest Boundary: A Comment on the Strategy for Regionalizing Evaporation

A. W. L. VEEN, R. W. A. HUTJES, W. KLAASSEN, B. KRUIJT & H. J. M. LANKREIJER
University of Groningen, Department of Physical Geography, P.O. Box 14, 9750 AA Haren, The Netherlands

ABSTRACT The regionalization problem in evaporation is a special case of the general problem of scale. A classification of scale is discussed which is based on the atmospheric stratification in Leaf Boundary, Internal Vegetation, and Surface Layers with successive decrease of decoupling from the Planetary Boundary Layer. The Penman-Monteith approach can be used to integrate lower level processes. For upscaling to the regional level we need to know more on the reaction of the Surface Layer to land surface discontinuities. This reaction is studied in the SHEAR project. First results indicate in a qualitative sense that due to the edge effects, many small forests may have somewhat higher evaporation loss than a single large forest.

INTRODUCTION

Predictions of climate change suffer from inadequacies of the General Circulation Models (GCM's). One of these is the coarse description of the exchange of latent and sensible heat at the land surface (Avissar & Verstraete, 1990). Important steps to tackle this problem have been made by the Hydrological Atmospheric Pilot Experiment (HAPEX), and similar experiments. The results show that a large forest (length scale of tens of kilometres) influences atmospheric circulation and precipitation (André et al., 1989). Modelling the regional atmospheric circulation and combining the mesoscale model with a GCM is the approach to solving this problem, but this method may be sensitive to small scale disturbances (André et al., 1989). On the other hand, over a region with small scale land surface variation (1-10 km) a Planetary Boundary Layer (PBL) develops, which is more or less uniform (Shuttleworth, 1988). In this situation one is tempted to simply average the fluxes. However, when internal boundary layers have developed this procedure is not without risks (André et al., 1990).

The problem of regionalizing evaporation, i.e. how to find in a simple way a correct regional value on the basis of point observations, is part of a more general problem of scale in the description of evaporation. Let us examine the scale problem somewhat closer.

A CLASSIFICATION OF SCALE IN EVAPORATION

Levels of scale in evaporation can be distinguished which are associated with the tangible elements stoma, leaf, plant, canopy and region (Jarvis & McNaughton, 1986; Baldocchi, 1989). Alternatively, one may envisage a
sequence of boundary layers, and link each scale level to one particular boundary layer.

The zero level of evaporation is given by the action of a single stoma. The gradient is the difference between concentration (or better: potential) of water vapour in the substomatal cavity and concentration at the leaf surface. The length scale of the gradient is of order $10^{-2}$ to $10^{-1}$ millimetre. Conductance, $g_{s0}$, depends on the cross sectional area of the stoma opening.

Water vapour leaving the stoma subsequently passes the Leaf Boundary Layer (LBL). Thus, at the first level of evaporation the total gradient is the sum of the gradients across the stoma and across the LBL. The latter has a length scale of the order of millimetres. Transpiration of a single leaf is described by an average leaf stomatal conductance ($g_{s1}$). If $g_{s1}$ is measured by porometry, then the much lower, parallel conductance of the cuticle is incorporated in the measured value of $g_{s1}$. Alternatively, the conductances at leaf level may be evaluated separately by modelling (Baldocchi, 1988).

At the second level of scale a second boundary layer, the Internal Vegetation Layer (IVL) should be distinguished, especially in tall vegetation. In forest with a distinct understorey, even two IVL's should be distinguished (cf. Black & Kelliher, 1989). This IVL is bounded at the "bottom" by the Leaf Boundary Layers. Its top could be chosen at the level $d + z_0$ (displacement height plus roughness length), at which level the downward extrapolation of the logarithmic profile of mean wind above the canopy reaches zero (Thom, 1971). In actual fact, exchange of air between the Internal Vegetation Layer and the air above the level $d + z_0$ does occur. This exchange is erratic and is associated with frequent counter gradient fluxes of heat and water vapour and needs a description other than conventional K-theory (Raupach, 1989). Still, a third segment having a length scale of metres is added to the total gradient. With this segment a separate average conductance, $g_{b2}$, is associated. At the IVL scale level the stomatal control is modified by the variation of illumination and temperature. In principle, stomatal conductance ($g_{s2}$) can be estimated by porometer observations of individual leaves.

At the third level of scale the total evaporation gradient is lengthened in the order of $10^1$ to $10^2$ metres by a third boundary layer, the atmospheric Surface Layer (SL), which is often defined as the turbulent, lowermost 10 % of the Planetary Boundary Layer (PBL). The conductance of the SL, $g_{b3}$, controls transport of latent heat towards the PBL. Canopy stomatal conductance, $g_{s3}$, exerts physiological control on evaporation by the action of the stomata of the leaves above the level of the IVL, i.e. above about 2/3 of canopy height.

At the fourth, regional level of scale the PBL acts as a sink for water vapour rising from the Surface Layer. Over land with small scale (1-10 km) changes in surface properties one (or more) internal boundary layers, caused by different upwind surfaces, may complicate the exchange of water vapour from land surface to the PBL overhead.

The picture which emerges from the preceding survey is this: the interaction between land surface and atmosphere proceeds through a succession of boundary layers, each being one or two orders of magnitude more extensive than the one at the lower level of scale, and each having a separate conductance. Therefore, there is a nested structure, in which each layer is decoupled to a varying degree from the conditions in the boundary layer at the next scale level. It can be shown (Jarvis & McNaughton, 1986) that, because of this decoupling - at each level of scale - a very small change in the associated stomatal conductances will
produce a very small change in evaporation, proportional to \((1-\Omega)\):

\[
dE/E = (1-\Omega) \frac{dg_s}{g_s}
\]  

(1)

where \(E\) is evaporation, \(\Omega\) the decoupling coefficient (0: no decoupling; 1: fully decoupled conditions), and \(g_s\) the stomatal surface conductance. In reality, a small change in the transpiration rate will immediately induce a change in the heat balance. Therefore, equation (1) does not describe the reaction of the evaporative system, which is described more fully by McNaughton (1988), and Black & Kelliher (1989). Still, it illustrates the importance of the decoupling phenomenon. Generally, the \(\Omega\) values at the successive levels of scale are not independent. Increasing wind speed in the SL above the canopy will tend to lower \(\Omega\) at lower levels. But stand architecture and leaf dimensions are bound to modify the relationship between the decoupling factors operative on IVL and LBL levels of scale.

The Penman-Monteith equation considers the canopy to be a "big leaf", characterized by two conductances. The aerodynamic conductance, \(g_a\), may be regarded as an effective conductance integrating the LBL, IVL and SL conductances. The surface conductance, \(g_s\), includes \(g_a\) and the stomatal action at lower levels (and any additional effects due to e.g. soil evaporation). It has been demonstrated that in some cases an estimate of bulk stomatal conductance derived from combined porometer observations of leaf stomatal conductances over the whole mass of leaves can be substituted for \(g_s\) (Dolman, 1988).

Using the conductances as defined by the big leaf model, the decoupling between canopy and Surface Layer is given by

\[
\Omega = \left[ 1 + \frac{g_a}{g_s} \right]^{-1}
\]  

(2)

where \(\varepsilon\) is \(s\lambda/c_p\), with \(s\) the change of slope of saturated specific humidity curve, \(\lambda\) the latent heat of vaporization of water and \(c_p\) the specific heat of air (McNaughton, 1988).

At the level of scale above the one where the Penman-Monteith equation has been applied so successfully, there is no approach available as yet, which equally elegantly integrates lower level i.e. within canopy processes. In the next section a strategy is outlined for regionalizing the Penman-Monteith approach, taking spatial differentiation into account.

Stomatal control is not operative when the vegetation is completely wet, and atmospheric control alone dominates evaporation during and shortly after rain. As in the dry situation decoupling complicates drying out at the levels of Leaf Boundary, Internal Vegetation, and Surface Layers. At the regional scale, canopies will not often be wet uniformly in space and time, adding to the spatial differentiation we have to cope with.

A STRATEGY FOR REGIONALIZING EVAPORATION

One possible strategy is to start with one-dimensional modelling, using the Penman-Monteith equation. The equation could be applied to each individual land unit. The second step would be to allow for the horizontal interactions between land units by correcting for advection in the Surface Layer. To do this three problems need to be solved. The first is how to obtain not only the characteristics of all types
of canopy, but also the values of the required meteorological variables above all canopy types. Secondly, one needs to know which part of a canopy conforms ideally to the Penman-Monteith equation, and thirdly how the fluxes near transitions are affected by advection from adjacent, different canopies.

Meteorological data are routinely collected over a grass surface. Some variables (net radiation, wind speed) can often be simply transformed to values appropriate to other surfaces but other variables (air temperature and humidity) need either to be measured - which is impractical - or modelled. In this respect, perhaps the relatively simple transformation scheme, proposed by McNaughton and Jarvis (1984) but not yet extensively tested, could be useful. For closed canopies the surface conductance can perhaps be approximated by bulk stomatal conductance, which - in principle - may be estimated from maximum specific stomatal conductance and Leaf Area Index, and empirical functions describing the feedbacks on soil moisture deficit and (modelled) meteorological conditions (Stewart, 1988). The available data on vegetation parameters such as albedo and canopy storage capacity for water of forest, are perhaps sufficient for making initial guesses.

Theoretical treatments of advection on the scale being considered here are numerous (for a review see Garratt, 1990). However, data on the effect of a simultaneous step change in surface roughness, displacement height and moisture availability on the Surface Layer are very scarce. Traditional rules of thumb for estimating the fetch indicate that air is in equilibrium with a new surface up to a height above that new surface of 1/300 to 1/100 of the fetch. In contrast, Gash (1986) concluded on the basis of measurements across a heath-forest interface that in his case height/fetch ratios were one order of magnitude larger. This means that as yet no proper description can be given of an every day situation like advection by wind blowing over grassland or over a low crop onto a forest.

Clearly, the strategy outlined above is not workable until more is known about Surface Layer processes near changes of land surface.

THE SHEAR PROJECT

The acronym SHEAR stands for Sleen Hydrometeorological Experiment on Advection and Regionalization. Sleen is the name of the Dutch village close to the experimental site. The aim of this experiment is to contribute to the knowledge of adjustment of the Surface Layer to land use changes, involving forest. The results should also give an idea to what extent one-dimensional modelling of forest water use is permitted for small forests. In the Netherlands this question has practical implications since the median size of forests is only 15 ha.

In the project the change in properties of air, passing from grassland to a mixed deciduous forest, is recorded. Wind speed, temperature and humidity is monitored above grass at a distance of 300 m from the forest edge, and at a number of levels above the forest canopy, at a distance of 200 m from the edge. The evaporative fluxes are measured intermittently at the same positions, using the eddy correlation technique. To obtain an impression of edge effects on water input to the forest floor, throughfall is recorded, and soil moisture is monitored. The change in air properties is being modelled using two approaches. One is to predict equilibrium values for air temperature, humidity and latent heat flux above the forest by transforming grassland
data according to the McNaughton and Jarvis (1984) scheme. The other approach is by modelling the air flow through and over the forest edge. Details of the methods, site and instrumentation are given by Kruijt & Van den Burg (1988).

OUTLINE OF FIRST RESULTS

The contrast between grass and forest with regard to temperature and humidity is illustrated in Fig. 1, which shows data from 23 to 27 May 1990, when the forest LAI was at about 2/3 of its maximum value. It was a period with unstable atmospheric conditions during the daytime, and very stable nights; no rainfall was recorded, and winds were moderate from N to NW directions resulting in fetches of 500-800 m both over the grass and the forest. Condensation is seen to lower the absolute humidity at night. Absolute humidity peaks in the morning are caused by the evaporation of dew.

The fluxes of latent and sensible heat (Fig. 2) show for this period a difference in energy partitioning: the Bowen ratio reaches about two over the forest, against around one over the grass. The sharp peaks in latent heat flux in the early mornings of the fourth and fifth day indicate the rapid evaporation of the dew formed at night. Similar peaks were recorded on other days as well, but they have been removed from the graph because the values were considered to be unreliable due to the likely presence of dew on the Krypton hygrometer.

![Temperature and absolute humidity graph](image)

**FIG. 1** Temperature and absolute humidity above grass and forest. X-axis show Julian day, 1990 (23-27 May).
Fig. 2 Fluxes of latent and sensible heat from grass and forest. X-axis shows Julian day, 1990 (23-27 May).

Fluxes measured at several levels above the forest during westerly winds (not shown here) demonstrate non-equilibrium conditions at a distance of 200 m from the forest edge. This subject is discussed in detail by Kruijt et al. (1991).

Fig. 3 shows forest and grass conductances. Aerodynamic conductance was derived directly from $u^*Z/u$, and surface conductance from subsequently solving the Penman-Monteith equation for $g_s$. The surface conductances do not differ very much. High - but not quite infinitely high - values of short duration occur early in the day when the vegetation was still wet because of the dew. Later in the day stomatal control of surface conductance reduces the values to normal levels. Aerodynamic conductance of the forest is seen to vary between three to ten times the value of $g_a$ for grass. The ratio of aerodynamic to surface conductance is variable. If rough daytime averages are considered, this ratio is about 30-50 for forest and 10-20 for grass.

Significant decoupling of the Internal Vegetation Layer is to be deduced from the profile of momentum flux in the forest (Fig. 4). The Surface Layer decoupling factors for grass and forest, $Q_g$ and $Q_f$ respectively, calculated from equation (2), are 0.2 for $Q_g$ and 0.06 for $Q_f$ around the middle of the day. As expected the forest is coupled more strongly to the Surface Layer than the grass. Both values, however, are lower than those given by Jarvis and McNaughton (1986). This could be consistent with the small scale heterogeneity of the landscape in the direction of prevailing winds, but a more complete analysis is needed.

The Surface Layer model (Klaassen, in prep.) takes the decoupling of the Internal Vegetation Layer into account, albeit implicitly, by allowing the wind to blow into the forest until the entire profile, i.e. both the IVL and the Surface Layer are adjusted. The "big leaf" model cannot describe the significant adjustment of the fluxes, which the
Surface Layer model predicts, because the former recognizes an above canopy Surface Layer only. It appears from the surface flow model that the edge effect is maximal when the canopy is wet ($g_s = \infty$), and rather less with average values for $g_s$. Under drought stress $g_s$ is very low and the edge effect is minimal. Adjustment occurs over a distance of about 10 times the forest height. For a Dutch forest of median size (15 ha) this means that evaporation may deviate from the estimate by the big leaf model, especially under conditions of ample water supply.

It is tempting to interpret the preliminary results of throughfall measurements (Fig. 5) as support for the validity of the models' prediction. However, further confirmation is necessary.
FIG. 4 Momentum fluxes in the forest, at three heights above the ground. X-axis shows Julian day, 1990.

FIG. 5 Weekly throughfall depths against distance to forest edge, measured with 60 gauges between 17/3 and 24/11/90.
DISCUSSION

There is a long way to go before it can be shown that the described strategy for regionalizing evaporation is feasible. In a study using HAPEX data Hutjes et al. (this volume), show that the transformation scheme may fail when horizontal distances are in the order of 10-50 km, and local conditions are disregarded. It has been demonstrated by measurements that the assumption of constancy of the flux in the Surface Layer does not hold near forest edges. This assumption may not be applicable at all in the Shuttleworth (1988) type "A" landscape with small scale variation. Therefore, Surface Layer modelling is necessary to analyse the problem of evaporation in the heterogeneous landscape more fully. From preliminary SL modelling it may be tentatively concluded that especially interception loss is increased by the presence of the forest edge. Therefore, regional evaporation may be higher in a landscape with many patches of forest (many edges) as compared with a landscape with the same total area of forest concentrated in large blocks (see Fig. 6).

![Modelled regional evaporation from a wet forest canopy.](image)

**FIG. 6** Modelled regional evaporation from a wet forest canopy. Y-axis shows evaporative flux with a hypothetical flux of available energy of 100 Wm$^{-2}$, and moderate wind. X-axis gives landscape periodicity (50% grass and 50% forest). E.g. 1 km represents alternation of 0.5 km grass and 0.5 km forest. * Evaporation without advection.

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