PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Forests and Regional-Scale Processes [and Discussion]

H. A. R. De Bruin, C. M. J. Jacobs, P. G. Jarvis, K. G. McNaughton, J. Milford, W. Kohsiek, P. Rowntree, J. L. Monteith, B. Gardiner and W. J. Shuttleworth

Phil. Trans. R. Soc. Lond. B 1989 **324**, 393-406
doi: 10.1098/rstb.1989.0054

Phil. Trans. R. Soc. Lond. B **324**, 393–406 (1989)
Printed in Great Britain

Forests and regional-scale processes

By H. A. R. de Bruin and C. M. J. Jacobs

Department of Physics and Meteorology, Duivendaal 2, 6701 *AP Wageningen, The Netherlands*

In this paper the surface energy balance of forests is considered on a regional scale. The interaction between the surface fluxes of sensible and latent heat on the one hand and the temperature (and humidity) of the planetary boundary layer (PBL) on the other is accounted for, as are entrainment processes at the top of the PBL. For this purpose, the one-dimensional PBL model of Troen & Mahrt is coupled to the simple 'single-leaf' vegetation model of Penman–Monteith. The surface conductance is described in accordance with Stewart, whereas the aerodynamic conductance is corrected for stability. The integrated model applies in stable, unstable and neutral conditions. Two forest types ('Thetford' and 'Les Landes') are modelled, and compared with grass and bare soil.

In summertime, calculated temperature, saturation deficit and depth of the PBL are greater over forest than over grass. The entrainment of warm, dry air at the top of the PBL appears to be important. It supplies the energy to maintain the evaporation of wet forest higher than net radiation. Well-known features, such as low transpiration and high evaporation rates (compared with grass) when the canopy is dry and wet respectively, are described correctly.

1. Introduction

Micrometeorological and hydrological research over the past few decades has provided a fairly complete picture of the peculiar micrometeorological behaviour of forests. A review on this topic is given by McNaughton & Jarvis (1983). Here we recall some important features.

1. Because of the large aerodynamic roughness, evaporation of intercepted water is usually higher for forest canopies than for grassland and arable crops.

2. Transpiration from forest is generally less than from short vegetation. This is related to the fact that the surface (or canopy) conductance of forests is usually smaller than that of short crops.

3. Forest has a relatively low albedo and thus absorbs more solar radiation than does short vegetation. As a result, net radiation of a forest can be about 20% higher in summer and 10% higher in winter than that of nearby grassland.

It is the purpose of this study to discuss this peculiar micrometeorological behaviour of forest on a regional scale, i.e. 10–50 km. In particular, we are interested in the surface energy balance (net radiation, sensible and latent heat flux, and soil heat flux) and in the momentum exchange.

At the scale of interest it is necessary to take into account the interaction between the surface fluxes of heat, water vapour and momentum on the one hand, and the properties (temperature, humidity and wind speed) of the overlying planetary boundary layer (PBL) on the other (Jarvis & McNaughton 1986). If, for instance, water vapour is brought into the atmosphere through evaporation at the surface, the humidity of the PBL will increase. This, in turn, will affect the surface water-vapour flux. Typically, the depth of the PBL is 1 km. This corresponds to a horizontal scale of 10–50 km, which is the scale of interest here.

Generally, the PBL is capped by an inversion. The air above the PBL is warmer and, usually, also drier. Because of turbulent vertical motion, generated by wind shear and surface heating, the height of the PBL increases during daytime. As a result, the air aloft is entrained into the PBL, tending to increase its (potential) temperature and decrease its (specific) humidity. Consequently, the entrainment process also affects the surface fluxes of heat and water vapour (de Bruin 1983; Brutsaert 1986; McNaughton & Spriggs 1986), and it is necessary to account for entrainment on a regional scale.

We consider the properties of the stable air above the PBL (e.g. the strength of the inversion) as independent boundary conditions and assume that the PBL does not affect these properties (except through entrainment). Moreover, we confine our calculations to flat, uniform surfaces. This allows us the use of a one-dimensional model.

In this study we use a coupled model, which consists of a number of sub-models adopted from literature. We designed the coupled model such that: (*a*) it is able to describe the stable and unstable PBL as well as the neutral PBL; and (*b*) it does not use too much computer time, so that it could possibly be implemented in, for example, models of climate.

For the PBL, we adopted the model of Troen & Mahrt (1986). At the surface the Penman–Monteith model is used. Moreover, the model of Stewart (1988) (see also Dolman *et al.* 1988) for the surface conductance of forest canopy is applied. For the stable surface layer, the findings of Holtslag & de Bruin (1988) are used, notably their description of the aerodynamic conductance, soil heat flux and net radiation. To illustrate the difference between forests and bare soils, we also use some results of Pan & Mahrt (1987) for bare soils.

Our model is designed to describe the surface energy balance and momentum exchange for forests in relation to the overlying PBL, including the entrainment process at its top. It is not meant to deal with micrometeorological processes taking place within the vegetation layer or in the soil. For that reason, these processes are parametrized rather crudely.

This study can be regarded as an extension of the work by de Bruin (1983) and McNaughton & Spriggs (1986), who only considered the well-mixed PBL. This study is also strongly related to the work by Pan & Mahrt (1987).

2. The model

(*a*) *General*

Our integrated model consists of a sub-model for the PBL developed by Troen & Mahrt (1986) and the Penman–Monteith 'single leaf' model for the vegetation and surface layer. For details of these sub-models the reader is referred to Troen & Mahrt (1986) and to Monteith (1981), respectively. Moreover, the papers by McNaughton & Jarvis (1983) and Jarvis & McNaughton (1986) are of relevance. Here we mention briefly only the most important features.

(*b*) *The* PBL*-submodel*

We first give a brief review of the different régimes into which the (idealized) PBL can be divided. In this we follow Holtslag & Nieuwstadt (1986).

Usually, the flow within the PBL is turbulent. Its turbulent structure can be described with three length scales: the height above the surface (z), the depth of the PBL (h) and the Obukhov length (L). The height, z, limits the size of the turbulent eddies to the ground and h limits the vertical extend of these eddies. The Obukhov length reflects the height at which the contribution to turbulent kinetic energy from buoyancy forces and from shear stress are comparable.

The *unstable* PBL ($L < 0$) can be divided into five distinct regimes (figure 1). The range $0.01 \leqslant z/h \leqslant 0.1$ is the *surface-* or *constant flux layer*, where the Monin–Obukhov similarity theory applies. At low wind speeds and high surface heating this layer transforms into the *free convection layer*. The *entrainment layer* is between about 0.8 z/h and 1.2 z/h. In addition, one can distinguish the *well-mixed layer* and the *near-neutral upper layer*; the latter layer can also exist under slightly stable conditions. The *stable* PBL can be divided into the *local scaling layer*, the *z-less scaling layer* and the *intermittency layer*, in addition to the surface layer.

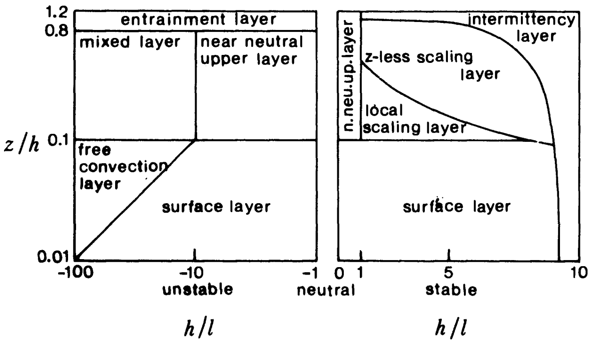

FIGURE 1. Definition of scaling regions in the PBL. (From Holtslag & Nieuwstadt 1986.)

At present, the behaviour of the surface layer, the well-mixed layer, the free convection layer, the local scaling layer and the z-less scaling layer is fairly well understood. The near-neutral upper layer and the entrainment layer are less well known, and at present there is no theory available to describe the intermittency layer. For more details, the reader is referred to Holtslag & Nieuwstadt (1986), and the cited literature.

The PBL model of Troen & Mahrt (1986), used in this study, contains most of our present knowledge of the PBL and has the following main characteristics.

1. It is one dimensional; in our version the model has 52 grid levels up to 4 km.

2. It is a first-order closure model, with a 'correction' term to allow an upward flux of heat and water vapour in the PBL when the vertical gradients of temperature and humidity vanish under convective conditions.

3. It describes the stable, unstable and the neutral PBL, and it provides a smooth transition between these states.

4. It behaves well under near-neutral conditions.

5. The depth of the PBL is determined by using a modified bulk-Richardson number.

6. When the PBL is well mixed the calculated vertical potential temperature gradient vanishes, but that for humidity does not.

(c) *The vegetation-surface-layer submodel*

The Penman–Monteith concept is used to describe the energy balance at the surface. It is applied to the layer between the ground and the first grid-point at height z_1 of the PBL model. (25 or 50 m). The vegetation is treated as if it were a single big leaf. The governing equations are listed in Appendix 1. It is common in hydrological and micrometeorological studies to approximate $q^*(\theta_0)$ (the saturation specific humidity at surface temperature (θ_0)) by $q^*(\theta_1)+s(\theta_0-\theta_1)$, which results in the well-known Penman–Monteith equation (for the symbols see the Appendix). However, for our application, where z_1 is at least 25 m, this approximation is too crude in some conditions. Therefore the basic equations listed in the Appendix are used and these are solved iteratively. (Note: the Penman–Monteith equation can lead to misinterpretation. For instance, according to this equation the latent heat flux density λE, tends to its so-called *equilibrium* value, $\lambda E_{eq} = (Q^*-G)\ s/(s+\gamma)$, when $r_a \to \infty$, whereas $\lambda E \to 0$, according to the equation (A 4), from which it is derived. The resolution of this paradox is that (Q^*-G) also tends to zero, because H vanishes when $r_a \to \infty$. This is not apparent from the Penman–Monteith equation on it own.)

Net radiation is described in accordance with de Bruin (1983) and Holtslag & Van Ulden (1983) for daytime conditions and in accordance with Holtslag & de Bruin (1988) for nighttime conditions. In stable conditions the expression proposed by Holtslag & de Bruin (1988) is used for the soil heat flux, whereas G is taken as proportional to Q^* during the day.

The aerodynamic resistances r_{av} and r_{ah} are described in accordance with Stewart & Thom (1973); in particular, their expressions for the roughness lengths for heat (z_T) and water vapour (z_q) are adopted. In the case of grass, z_T is taken equal to z_q and $\ln(z_0/z_T) = 2$, where z_0 is the roughness length for momentum. Between z_0 and z_1 the aerodynamic resistances for heat and water vapour are assumed equal and evaluated with the bulk-Richardson number approach of Louis (1979) in unstable conditions. This approach yields almost the same results as the Dyer–Hicks stability functions (Dyer 1974) as used, for example, by McNaughton & Spriggs (1986). In stable conditions the stability correction proposed by Holtslag & de Bruin (1988) is applied. The surface resistance, r_c, for dry forests is described in accordance with Stewart (1988), taking account of the work by Stewart & de Bruin (1985) and Dolman *et al.* (1988).

(*d*) *Coupling and initial and boundary conditions*

The integrated model requires the vertical profiles of temperature, humidity and wind velocity up to 4 km as initial conditions (at $t = 0$). There is also a set of parameters such as roughness length, zero-plane displacement, albedo and surface emissivity that must be specified. At all times, t, the model needs the global radiation ($K\downarrow$), geostrophic wind speed and cloud cover (clouds above the PBL) as 'driving forces'. In most of the simulations global radiation is calculated from cloud cover, etc. in accordance with Holtslag & Van Ulden (1983).

In the first time-step the surface fluxes of heat, water vapour and momentum are computed from the initial temperature, humidity and wind profiles. These are used as input to the model of the PBL to evaluate the new profiles. This allows the calculation of the surface fluxes for the next time-step, taking into account the new value of the global radiation, etc.

(*e*) *Model verification*

Complete sets of micrometeorological data for both the surface energy balance and the PBL are scarce. For model verification we have used the data from the KNMI tower site at Cabauw in The Netherlands (Driedonks 1981). This data set is confined to the well-mixed PBL. It includes information on global and net radiation, sensible and latent heat flux, momentum flux,

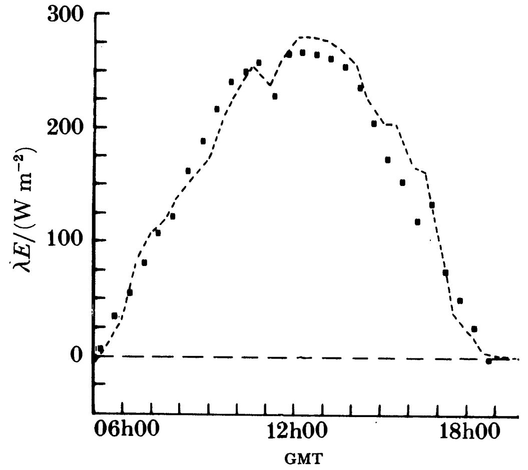

FIGURE 2. Comparison of the calculated latent heat flux density λE(—) with the observed one (■) on 5 August 1977. The observed global radiation is used as input.

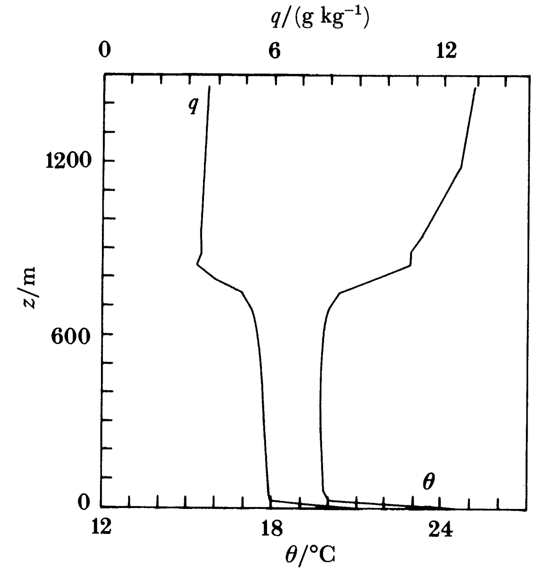

FIGURE 3. Calculated profiles of the potential temperature (θ) and the specific humidity (q) for 5 August 1977, 12h00 GMT.

the depth of the PBL and the vertical profiles of potential temperature (θ), specific humidity (q) and wind.

The site is primarily surrounded by grassland. We use its values of r_c derived for the same data set by McNaughton & Spriggs (1986). The present model describes the Cabauw data set fairly well and we obtain results similar to those of de Bruin (1983) and McNaughton & Spriggs (1986), who used a well-mixed model of the PBL that was similar to that of Driedonks (1981, 1982). As an example we present a comparison between the measured water-vapour flux and that determined with the present model for 5 August 1977 (figure 2). The calculated profiles of the potential temperature and the specific humidity for 12h00 Greenwich Mean Time (GMT) are also shown (figure 3) to illustrate that θ is fairly well mixed, whereas q is not.

3. Model simulations

(a) Dry canopy

Because of the lack of relevant data for the PBL, we use the profiles of temperature, humidity and wind speed observed at Cabauw on 5 August 1977, 05h00 GMT as initial conditions for the simulation of the energy-balance and characteristics of the PBL for forests with a dry canopy. At the surface boundary conditions are taken typical of two forest types: (1) a 'Thetford-type' and (2) a 'Les Landes-type'. For 'Thetford', the surface conductance relations proposed by Dolman *et al.* (1988) are used. For 'Les Landes', use was made of recent observations made during the HAPEX field experiment (Gash *et al.* 1988). These two forests have rather different surface conductances, see Shuttleworth (this symposium). As a result, the Bowen ratio over Les Landes seldom exceeds 1.5, whereas over Thetford it often reaches values of 2 or 3 (Stewart & Thom 1973).

Roughness-length, z_0, and zero-plane displacement, d, are taken from de Bruin & Moore (1985) for 'Thetford' and from Gash *et al.* (1988) for 'Les Landes'. The albedo is taken as 0.1 for both forests.

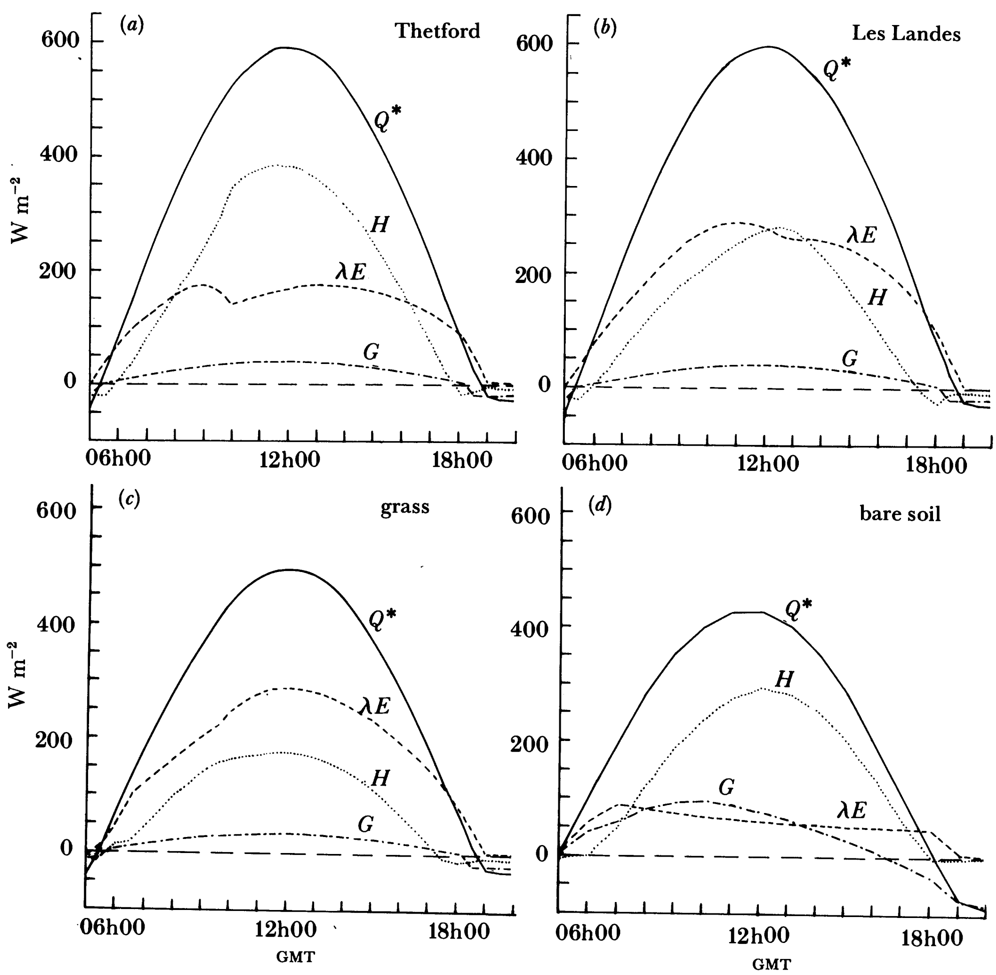

FIGURE 4 *a–d*. Modelled evolution of the components of the energy balance: net radiation, Q^* (—); latent heat flux density, λE(- - -); sensible heat flux density, H(····); and soil heat flux density, G (–·–). (*a*) 'Thetford'. (*b*) 'Les Landes'. (*c*) Grass. (*d*) Bare soil (sandy clay loam).

Figures 4*a*, *b* show the calculated components of the energy balance for these two forests, whereas figures 5–7 show the evolution of the depth of the PBL, the mean potential temperature of the PBL and the saturation deficit 2 m above the vegetation, respectively. Note that during the day the PBL is almost well mixed under these (convective) conditions. For comparison, the results for grassland are also given in figures 4–7 and some results for bare soil (sandy clay loam) as evaluated by Pan & Mahrt (1987) are presented in figures 4 and 5.

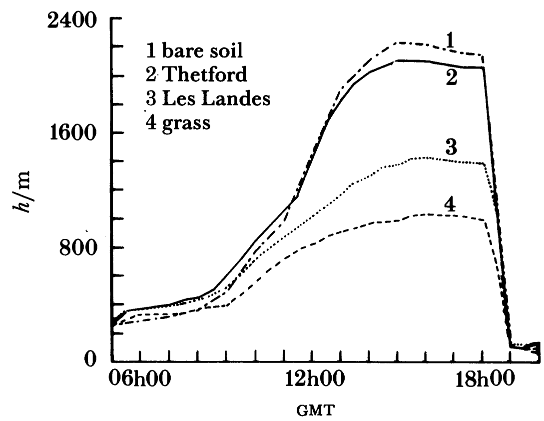

FIGURE 5. Modelled evolution of the depth, h, of the PBL over four surface types: 'Thetford' (—); 'Les Landes' (·····); grass (- - -); and bare soil (sandy clay loam) (–·–).

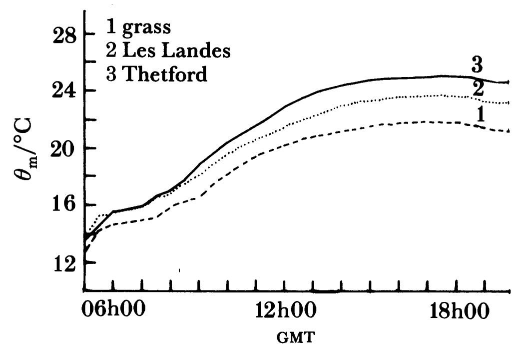

FIGURE 6. Modelled evolution of the mean potential temperature, θ_m, of the PBL over three types of vegetation: 'Thetford' (—); 'Les Landes' (·····); and grass (- - -).

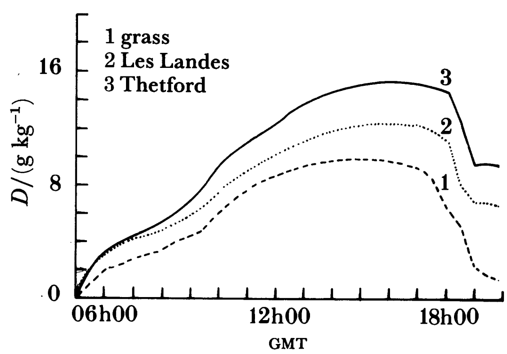

FIGURE 7. Modelled evolution of the saturation deficit, D, at 2 m above vegetation: 'Thetford' (—); 'Les Landes' (·····); and grass (- - -).

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The transpiration and sensible heat flux are significantly different for the different vegetation types because of the different environmental responses of their surface resistance. Grass has a somewhat larger transpiration than 'Les Landes', which in turn transpires significantly more than 'Thetford'. The calculated net radiation for forests is almost 20% higher than that for grass during the day. As a result, the sensible heat flux over grass is smaller than over 'Les Landes' and much smaller than over 'Thetford'. Because surface heating is the dominant source in the generation of turbulence, the depth of the PBL over 'Thetford' becomes greater than over 'Les Landes' and grass. This is important in air pollution problems, for example. Potential temperature and saturation deficit also reach their highest values over 'Thetford'.

The transpiration curve for 'Thetford' shows an interesting feature. It rises rapidly in the early morning, until about 08h00 GMT, after which it declines and then stays almost constant between about 11h00 and 16h00 GMT. This is a result of the particular relation between the surface resistance, r_c, and the saturation deficit, D. In the early morning D is small as, correspondingly, is r_c. After 08h00 GMT the PBL grows rapidly, resulting in the entrainment of dry, warm air into the PBL. This leads to an increase in D and consequently to a rise in r_c, giving a decline of λE in this case. This feature has been reported in the literature (McNaughton & Jarvis, 1983). Note that the transpiration curve for 'Les Landes' does not show a dip. The dip also disappears for 'Thetford' in the case of moister and cooler air above the PBL (not shown here).

In general, the calculated values for Q^*, λE and H are in agreement with observations made at Thetford and Les Landes under the convective conditions considered.

(b) *Wet canopy*

In the case of wet canopies, the evaporation from forests can be as large as, or even larger than, the water equivalent of net radiation (Stewart 1977; McNaughton & Jarvis 1983). Simulations of wet forest are hindered by the lack of data of the PBL for rainy days. A simulation was made by using constructed initial temperature, humidity and wind profiles, typical for wet conditions, and an assumed completely overcast sky. The results of this simulation (for 'Thetford') are shown in figure 8. For comparison, net radiation and λE for the assumed dry canopy in the same conditions are also given.

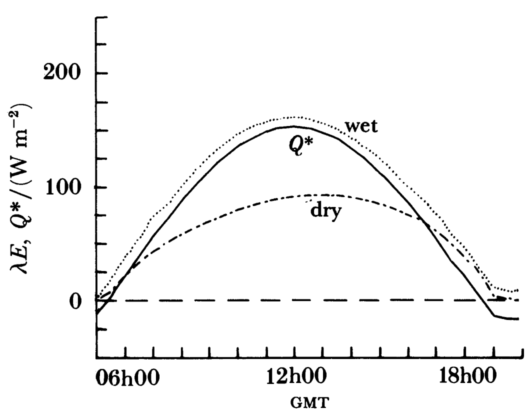

FIGURE 8. Modelled net radiation (Q^*) and latent heat flux density (λE) for 'Thetford' with a completely overcast sky and constructed initial profiles: Q^*, wet canopy (—); λE, wet canopy (·····); λE, dry canopy (-·-).

It can be seen that the calculated evaporation of the wet canopy does indeed exceed net radiation (by about 10 W m^{-2}) the entire day. (Note: in reality, all the intercepted water would be consumed after a certain time once precipitation ceased.)

In the literature (Stewart 1977; McNaughton & Jarvis 1983), the question has arisen as to where the additional energy comes from to maintain λE higher than Q^*. According to our calculations, the mean temperature of the PBL remains almost constant. This implies that the entrained warm air coming from above the PBL supplies the required energy. Note that in this case the calculated wind speed at z_1 was about 4 m s^{-1}. In fact, it was found that λE increases with increasing wind speed. The issue of evaporation from forest with a wet canopy merits further study.

(c) *Drying canopy, onset of deep convection*

Figure 9 shows the results of a simulation in which a wet 'Thetford' canopy in the morning (until 09h00 GMT), dries until 12h00 GMT and then remains dry. The results are as expected: in the morning λE exceeds Q^* but falls below Q^* after drying phase.

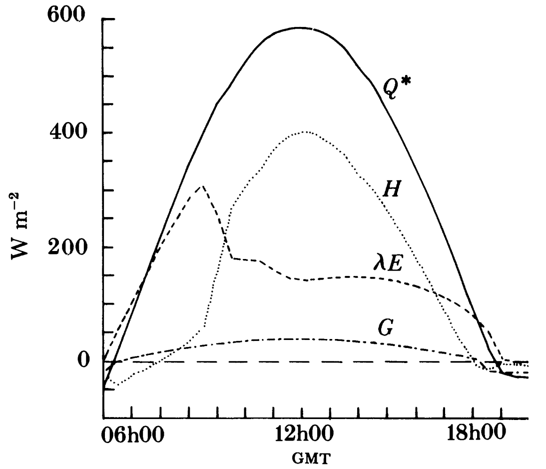

FIGURE 9. Modelled evolution of energy-balance components for a drying 'Thetford-type' forest: net radiation, Q^* (—); latent heat flux density, λE (---); sensible heat flux density, H (·····); and soil heat flux density, G (-·-·).

This case has been selected to draw attention to an important feature, namely the probable onset of deep convection that can lead to showers. This process occurs when the atmosphere is in a conditionally or latent unstable state. Moreover, relatively moist surface air is required and the (integrated) sensible heat flux at the surface must exceed a certain threshold value (see, for example, McIntosh & Thom 1978). This implies that the case we consider here favours the onset of deep convection. In the morning the lower atmosphere becomes moist, whereas after the drying phase the surface heating becomes high. This suggests that over forests the hydrological cycle (as far as this is related to deep convection) turns over faster than over grassland, which does not show this peculiar behaviour. This feature possibly explains the rainfall maximum that occurs over a forested area in the centre of The Netherlands, and it may be of importance for tropical forests (see Shuttleworth, this symposium). This phenomenon also merits further study.

(*d*) *Advection*

Often forests are so small that horizontal advection plays an important role and the properties of the upwind terrain then affect the forest's energy balance. We considered the transition from grassland to forest, discussed by McNaughton & Jarvis (1986), but also investigated the reverse case. Again, the profiles of the PBL at Cabauw on 5 August 1977 were used to start with. Moreover, a trajectory approach has been applied in which a column of air was

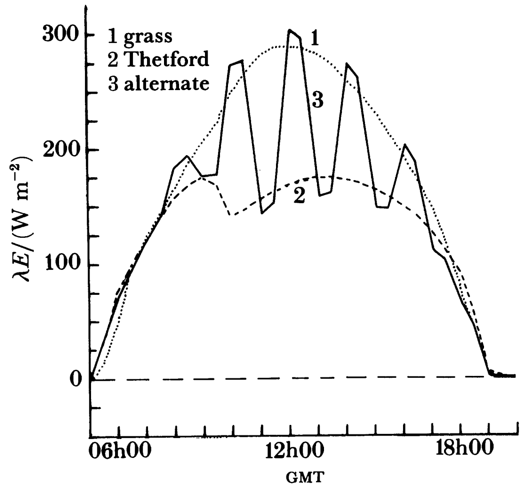

FIGURE 10. Modelled latent heat flux density λE for alternately grassland and 'Thetford' (—), uniform grassland (·····), and uniform 'Thetford' (- - -).

imagined to pass over dry grass and dry forest alternately. It was assumed that the air passed over forest until 08h00 GMT, then for the next hour over grass, for the next hour over forest, and so on. (Note: the net radiation for grass and forest is different.) Figure 10 shows the calculated transpiration in this case, together with the transpiration calculated for uniform grass and uniform forest (previously shown in figure 4). The upwind terrain is clearly seen to influence transpiration. The transpiration from grassland is enhanced by an upwind forest, that of forest is suppressed by upwind grassland, in this case.

4. Discussion and concluding remarks

In this study a coupled PBL-vegetation model has been used to describe the micrometeorological behaviour of forests on a regional scale. The model realistically simulates evaporation and transpiration from forest with wet or dry canopy. It appears that the entrainment of warm and dry air at the top of the PBL plays a significant role. In the case of a wet forest canopy, it supplies the energy to allow evaporation in excess of net radiation.

In summertime the PBL grows much more rapidly over forest and reaches greater heights than over grassland. The PBL over a Thetford-type forest becomes almost as high as over bare soil. The temperature and atmospheric saturation deficit over forests also become higher than over grass.

It appears that the relation between surface conductance and environmental parameters plays a crucial role. The surface conductances of the two forest types considered here are significantly different. If, as suggested by McNaughton & Jarvis (1983), the response of surface conductance to environmental variables changes from one forest to the next, this would significantly hamper the modelling of the energy balance of forests. However, on the basis of evidence presented at this meeting (see paper by Shuttleworth) it may simply be that one of the forests used in our simulation studies is atypical.

We gratefully acknowledge Professor Hua-Lu Pan and Dr Bert Holtslag for their advice and for making available the computer code of the sub-model of the PBL. Dr Kenneth Mitchell (AFGL, Cambridge, Massachusetts, U.S.A.) is thanked for giving us permission to use this model. Also, we are grateful to Dr Leo Kroon and Bart van den Hurk for their helpful comments. This study was sponsored to a large extent by the European Community under contract number EV4C/0015NL.

References

Brutsaert, W. 1986 Catchment-scale evaporation and the atmospheric boundary layer. *Water Resour. Res.* **22**, 39–45.

De Bruin, H. A. R. 1983 A model for the Priestley–Taylor parameter α. *J. Climate appl. Met.* **22**, 572–578.

De Bruin, H. A. R. & Moore, C. J. 1985 Zero-plane displacement and roughness length for tall vegetation, derived from a simple mass conservation hypothesis. *Boundary-Layer Met.* **31**, 39–49.

Dolman, A. J., Stewart, J. B. & Cooper, J. D. 1988 Predicting forest transpiration from climatological data. *Agric. Forest Met.* **42**, 339–353.

Driedonks, A. G. M. 1981 Dynamics of the well-mixed atmospheric boundary layer. *Scientific report W.R. 81–2.* (189 pages.) De Bilt: Royal Netherlands Meteorological Institute.

Driedonks, A. G. M. 1982 Models and observations of the growth of the atmospheric boundary layer. *Boundary-Layer Met.* **23**, 283–306.

Dyer, A. J. 1974 A review of flux-profile relationships. *Boundary-Layer Met.* **7**, 363–372.

Gash, J. H. C., Shuttleworth, W. J., Lloyd, C. R., André, J. C., Goutorbe, J. P. & Gelpe, J. 1988 Micrometeorological measurements in Les Landes forest during HAPEX–MOBILHY. *Agric. Forest Met.* (In the press.)

Holtslag, A. A. M. & de Bruin, H. A. R. 1988 Applied modelling of the nighttime surface energy balance over land. *J. Climate appl. Met.* **27**, 689–704.

Holtslag, A. A. M. & Nieuwstadt, F. T. M. 1986 Scaling the atmospheric boundary layer. *Boundary-layer Met.* **36**, 201–209.

Holtslag, A. A. M., Van Ulden A. P. 1983 A simple scheme for daytime estimates of the surface fluxes from routine weather data. *J. Climate app. Met.* **22**, 517–529.

Jarvis, P. G. & McNaughton, K. G. 1986 Stomatal control of transpiration: scaling up from leaf to region. *Adv. ecol. Res.* **15**, 1–49.

Louis, J.-F. 1979 A parametric model of vertical eddy fluxes in the atmosphere. *Boundary-Layer Met.* **17**, 187–202.

McIntosh, D. H. & Thom, A. S. 1978 *Essentials of meteorology*, 4th edn. (240 pages.) London: Wykeham.

McNaughton, K. G. & Jarvis, P. G. 1983 Predicting effects of vegetation changes on transpiration and evaporation. In *Water deficits and plant growth* (ed. T. T. Kozlowski), vol. 7, pp. 1–47. New York: Academic Press.

McNaughton, K. G. & Spriggs, T. W. 1986 A mixed layer model for regional evaporation. *Boundary-Layer Met.* **34**, 243–262.

Monteith, J. L. 1981 Evaporation and surface temperature. *Q. Jl R. met. Soc.* **107**, 1–27.

Pan, H.-L. & Mahrt, L. 1987 Interaction between soil hydrology and boundary-layer development. *Boundary-Layer Met.* **38**, 185–202.

Stewart, J. B. 1977 Evaporation from the wet canopy of a pine forest. *Water Resour. Res.* **13**, 915–921.

Stewart, J. B. 1988 Modelling surface conductance of pine forest. *Agric. Forest Met.* **43**, 19–35.

Stewart, J. B. & de Bruin, H. A. R. 1985 Preliminary study of dependence of surface conductance of Thetford forest on environmental conditions. In *The forest–atmosphere interaction* (ed. B. A. Hutchinson & B. B. Hicks), pp. 91–104. Dordrecht: D. Reidel.

Stewart, J. B. & Thom, A. S. 1973 Energy budgets in pine forest. *Q. Jl R. met. Soc.* **99**, 154–170.

Troen, I. B. & Mahrt, L. 1986 A simple model of the atmospheric boundary layer; sensitivity to surface evaporation. *Boundary-Layer Met.* **37**, 129–148.

Appendix 1

At the surface, the available energy (Q^*) is partitioned between sensible heat flux density, H, latent heat flux density, λE, and soil heat flux density, G; other terms (e.g. biological production) are neglected:

$$Q^* = H + \lambda E + G, \tag{A 1}$$

in which Q^*, H, λE and G are calculated from:

$$Q^* = (1-r)\, K\!\downarrow - \epsilon_0 \sigma T_0^4 + \epsilon_1 \sigma T_1^4, \tag{A 2}$$

$$H = \rho_a c_p (\theta_0 - \theta_1 / r_{ah}), \tag{A 3}$$

$$\lambda E = \rho_a \lambda[(q^*(\theta_0) - q_1)/(r_{av} + r_c)], \tag{A 4}$$

$$G = c_G Q^* \quad \text{(unstable)}, \tag{A 5}$$

and

$$G = c_H H + c_Q Q_1^* \quad \text{(stable)}. \tag{A 6}$$

List of symbols used in equations

Q^*	net radiation ($\mathrm{W\,m^{-2}}$)
H	sensible heat flux density ($\mathrm{W\,m^{-2}}$)
λE	latent heat flux density ($\mathrm{W\,m^{-2}}$)
λ	latent heat of vaporization of water ($\mathrm{J\,kg^{-1}}$)
G	soil heat flux density ($\mathrm{W\,m^{-2}}$)
r	albedo (dimensionless)
$K\!\downarrow$	global (or solar) radiation ($\mathrm{W\,m^{-2}}$)
ϵ_0	emissivity of the Earth's surface (dimensionless)
ϵ_1	apparent emissivity of the sky, including clouds (dimensionless)
σ	Stefan–Boltzman constant ($5.6705 \times 10^{-8}\ \mathrm{W\,m^{-2}\,K^{-4}}$)
T	temperature (K)
ρ_a	air density ($\mathrm{kg\,m^{-3}}$)
c_p	specific heat of air at constant pressure ($\mathrm{J\,kg^{-1}\,K^{-1}}$)
θ	potential temperature (K)
r_{ah}	aerodynamic resistance for sensible heat ($\mathrm{s\,m^{-1}}$)
r_{av}	aerodynamic resistance for water vapour ($\mathrm{s\,m^{-1}}$)
q	specific humidity ($\mathrm{kg\,kg^{-1}}$)
$q^*(\theta)$	saturation specific humidity at θ ($\mathrm{kg\,kg^{-1}}$)
s	slope of saturation humidity–temperature curve at θ_1 ($\mathrm{K^{-1}}$)
D	saturation deficit ($= q^*(\theta) - q$)
r_c	canopy, or surface, resistance ($\mathrm{s\,m^{-1}}$)
c_G	G/Q^*, taken here as 0.07
c_H	constant of proportionality, taken as $-\frac{1}{4}$
c_Q	constant of proportionality, taken as $+\frac{1}{3}$
Q_1^*	$-\sigma T_1^4(\epsilon_0 - \epsilon_1)$, the net isothermal long-wave radiation ($\mathrm{W\,m^{-2}}$)
0, 1	subscripted indices for surface and first level of the model for the PBL, respectively

Discussion

P. G. Jarvis (*Department of Forestry and Natural Resources, University of Edinburgh, U.K.*). Would Dr de Bruin please elaborate on the feature of the model that results in very different growth in height of the PBL over different surfaces? What feature of the model gives growth to *ca.* 2.5 km over both bare soil and Thetford, and half that height over grass?

H. A. R. DE BRUIN. The depth of a well-mixed PBL at a particular time is roughly proportional to the square root of the time integral of the sensible heat flux density integrated from about sunrise to that time. Because the surface heat fluxes for the four surfaces considered are significantly different, the resulting growth of the PBLS necessarily show significant differences.

K. G. MCNAUGHTON (*Plant Physiology Division, DSIR, Palmerston North, New Zealand*). In experiments with our simple mixed-layer model, Tom Spriggs and I found that the surface fluxes take something like 1 h to adjust to a step change in net radiation or canopy resistance. This is about the time it takes for a new saturation deficit to become established throughout the PBL. This means that a column of air must travel something like 20 km over a 'new' surface before it is fully adjusted, and local advection with the convective PBL must be significant for any forests smaller than that. Would Dr de Bruin comment on the time that it takes his model to adjust to such changes during daytime (convective) conditions?

H. A. R. DE BRUIN. Our calculated surface fluxes change sharply after a step-change at the surface. It appears that 'history' plays a dominant role in the question of how, or how fast, the PBL and the surface fluxes adjust to new surface conditions. In the case of a transition from grass to forest, the air adjusts after about 2 h, and then has the same properties that it had over uniform forest. However, after a transition from forest to grass an equilibrium state is established that differs significantly from that which would occur over uniform grass. The explanation for this is that the PBL grows much higher over forest than over grass, and, after a forest–grass transition, stays much higher than it would over uniform grass. Under the convective conditions considered here there is no mechanism to bring the PBL back to its grass-equilibrium condition.

J. MILFORD (*Department of Meteorology, University of Reading, U.K.*). The model runs over different surfaces all started from a similar atmospheric state: are there not significant differences in the initial conditions due to the different fluxes during the previous night?

H. A. R. DE BRUIN. For convective conditions, to which our examples refer, it can be shown that the surface fluxes are not very sensitive to the initial conditions. Generally, however, initial conditions are important.

W. KOHSIEK (*Koninklijk Nederlands Meteorologisch Instituut, De Bilt, The Netherlands*). I have three questions.

1. Are the model's surface fluxes sensitive to $\Delta\theta_v$, i.e. the jump of the virtual temperature at the inversion?
2. Are there experimental verifications of this model over forest?
3. How fast does a new equilibrium boundary layer grow if the air is flowing across a discrete change in surface characteristics (e.g. grass–forest)?

H. A. R. DE BRUIN. I shall answer Dr Kohsiek's questions separately.

1. In the present PBL model, $\Delta\theta_v$ is not an explicit model parameter. Earlier work based on the well-mixed PBL model by Tennekes and Driedonks revealed the surface fluxes not to be very

sensitive to variation in $\Delta\theta_v$. The depth of the PBL, on the other hand, appears to be more sensitive.

2. The problem in verifying any PBL model is lack of experimental data, and this also applies to our model; but the calculated surface fluxes are consistent with measurements.

3. The answer to Dr Kohsiek's third question is similar to that I gave to Dr K. McNaughton.

P. Rowntree (*Meteorological Office, Bracknell, U.K.*). In changing from grass to forest in the calculations, was the albedo changed as well as the resistances? If so, can Dr de Bruin separate the roles of these two: that is, the role of changing the total energy available, in contrast to changing the partition between sensible and latent heat fluxes?

H. A. R. de Bruin. We have not yet done systematic sensitivity studies, but we do know that changes in the partition between sensible and latent heat are primarily due to differences in the surface resistances, whereas differences in net radiation are mainly caused by changes in albedo.

J. L. Monteith, F.R.S. (*International Crops Research Institute for the Semi-Arid Tropics, Hyderabad, India*). How typical is Thetford forest?

P. G. Jarvis (*Department of Forestry and Natural Resources, University of Edinburgh, U.K.*). Perhaps I could answer this question. Thetford forest is not typical of most of the U.K. forest. At Thetford there are about 700 trees per hectare and the leaf area index is only about 3. There are many gaps between trees where the crowns do not touch. This is very different from the typical lodgepole pine and spruce forests elsewhere in the U.K., which have 1000–2500 stems per hectare and a leaf area index of 6–12 (1 hectare = 10^4 m^2).

B. Gardiner (*Forestry Commission Northern Research Station, Roslin, U.K.*). I should like to pick up on the suggestion that came from Dr de Bruin's work that a drying forest might be an initiator of deep convection. In a convective cloud study in Montana on days when there was no obvious mesoscale forcing there was evidence that convection was initiated over small plantations of trees in an area that was generally grass covered. This observation seems to support Dr de Bruin's results.

W. J. Shuttleworth (*NERC Institute of Hydrology, Wallingford, U.K.*). This is valuable, independent confirmation of a similar phenomenon observed over the Landes forest during the HAPEX experiment in southwest France, and is discussed in greater detail by Dr André in the next paper.