Parameterising the Evapotranspiration and Energy Balance of a Natural Eucalypt Forest

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Abstract: This paper describes the parameterisation of transpiration and soil evaporation in a catchment scale model that couples soil moisture storage, surface energy balance and the atmospheric boundary layer. The model determines the energy balance at a canopy surface and the soil surface by coupling bulk stomatal resistance and soil surface vapour pressure to the soil moisture content. Energy flux data from a natural eucalypt forest during spring and summer were used to test parameterisations of canopy conductance and vapour pressure at the soil surface. Some existing parameterisations are found to be adequate, whereas others were deemed to require modification for this environment. The data also indicated that heat storage in the deep canopy was a substantial proportion of the energy budget at the beginning and end of the day, with heat stored in the trunks the major component of it. We modelled this using the "force-restore" method widely used for soil heat flux which proved adequate and is relatively simple to do.

Keywords: Forest; Energy balance; Canopy conductance

1 INTRODUCTION

Hydrology has progressed to the stage where the atmosphere—surface interaction can be incorporated into catchment models for hydrological research and as a management tool. For large areas to be well managed we need better methods of measurement of forest energy balance than we currently have. There have been few successful attempts at modelling these surface—atmosphere interactions at catchment scale [e.g. Sellers and Lockwood, 1981], and the problem is complicated by the different time and space scales of the processes. Braud et al., [1995] contend that soil-vegetation-atmosphere models in the literature tend to display an imbalance in detail of their various compartments according to the special interests of their authors. However, this apparent imbalance may also reflect the difference in spatial and temporal scales that need to be accommodated by the models to adequately describe the processes simulated, and that are of interest to land managers. Besides, validation of models that couple atmospheric, surface and soil moisture processes is problematic, and perhaps impossible [Oreskes et al., 1994].

In the south-west of Western Australia (W.A.), water supply catchments are largely forested with natural dry sclerophyll eucalypt forest, that produce runoff at the rate of ~1-10% of rainfall. Some of these catchments are subject to bauxite mining, mine rehabilitation, and forestry logging. Modelling (supplemented with field experiments) presents the only feasible method for assessing the potential for risk to the water resources on a large scale. With evapotranspiration accounting for over 90% of rainfall in these catchments, a small change in the energy balance partitioning between soil evaporation and transpiration may have relatively large consequences for runoff.

This paper presents the analysis of field data to parameterise a catchment energy and water balance model. Particular attention is given to identifying the controls on canopy conductance, and simulating soil surface vapour pressure which drives soil evaporation. Getting these right makes a huge difference to the partitioning between evaporation at the soil and canopy surfaces. We also present a new model of canopy heat storage.

2 MODEL DESCRIPTION

The model COUPLE attempts to bridge the gap between the temporal and spatial scales of catchment-atmosphere interactions by using a small (~1 km^2) catchment scale model for water balance and runoff representation based on three conceptual soil moisture stores running at daily time steps, and a more detailed physically based canopy and soil surface energy balance model run at hourly or sub-hourly time steps, with a simpler (and larger spatial scale, ~10 km) convective atmospheric boundary layer model at sub-hourly time steps. Thus each stage of the catchment-surface-atmosphere interaction is simulated without excessive computation or detail where it is not required. Most details are described elsewhere [Silberstein and Sivapalan, 1996; Silberstein et al., in press] and a brief description is given here of a few aspects of the energy balance.

2.1 The Water Balance Submodel

The water balance component of the model centres around three lumped conceptual soil moisture stores on a catchment scale, referred to as the A, B, and F stores. The A store is the shallow seasonal perched aquifer system; the B store is the deep permanent groundwater; and the F store is an intermediate unsaturated zone, which has a high degree of seasonality, especially under forest. The F store is the major water source for the forest during the summer drought (~6 months), with roots penetrating throughout its depth (to ~40 m). The dynamics of the stores, through the redistribution of water within the profile, controls the short term runoff response, the long-term water balance, and has a major influence on the short term energy balance, through transpiration and soil surface evaporation.

2.2 The Energy Balance Submodel

The atmosphere-surface interaction is driven by the net absorption of radiation at two conceptual surfaces, (vegetation and soil) and the partitioning of it into turbulent fluxes to the atmosphere, and surface heat storage. The energy balance is based on Choudhury and Monteith [1988] and Choudhury [1989], with some modifications [Silberstein and Sivapalan, 1996] and the inclusion of canopy heat storage. We neglect advection and assume that for a mature natural forest, with a stable total biomass, net photosynthesis is also negligible. The energy balance at the soil and vegetation surfaces, respectively, can be written:

$$R_{ns} = H_s + \lambda E_s + G \qquad (1)$$
$$\text{and} \quad R_{nv} = H_v + \lambda E_v + J \qquad (2)$$

with R_n the net radiation at the effective "surface", λ is latent heat of vaporisation of water, E is the mass water vapour flux density, H is the sensible heat flux, G the heat conduction into the soil, J the change in heat storage of the canopy, and the subscripts s, and v, refer to the soil and canopy surfaces, respectively. H and E are modelled using gradient diffusion theory and the electrical resistance circuit analogy with bare soil evaporation acting in parallel with the transpiration and evaporation of intercepted rainfall into a common canopy air layer. J is often neglected as in many situations it is minor in comparison to the other terms. However, for a forest with a high canopy enclosing large volumes of air and a high woody biomass, the canopy heat storage can be significant, particularly near dawn and dusk [Silberstein et al., 2001].

2.3 Evaporation from the Soil Surface

Evaporation from the soil surface, E_s, is dependent on a soil surface-canopy air vapour pressure gradient:

$$\lambda E_s = \rho C_p/\gamma\ (e_s - e_b)/r_w \qquad (3)$$

where ρ is the density of moist air; γ is the psychrometric constant; C_p is the specific heat of air; e_b and e_s are the water vapour pressure in the canopy air and immediately above the soil surface, respectively; and r_w is the resistance to turbulent diffusion from the soil surface to canopy air for latent and sensible heat [Choudhury, 1989]. Often [Choudhury, 1989; Silberstein et al., 1999] the soil surface vapour pressure is modelled [Philip, 1957]:

$$e_s = e_{sat}(T_{ss})\ exp(g\psi_s/R_w T_{ss}) \qquad (4)$$

where T_{ss} is temperature of the soil surface, R_w is the gas constant for water vapour, g is acceleration due to gravity, $e_{sat}(T_{ss})$ saturation vapour pressure at soil temperature T_{ss}, and ψ_s is the moisture potential of the surface soil, dependent on soil moisture content of the shallow soil and given by:

$$\psi_s = \psi_o\ (\theta/\theta_{sat})^{-\xi} \qquad (5)$$

with ψ_o representing the capillary fringe; θ the soil moisture content, and θ_{sat} the saturation moisture storage; and ξ is set at measured or literature values for this representation of the soil moisture characteristic. However, (4) calculates the vapour pressure in the pore spaces and overestimates e_s, and Lee and Pielke [1992] suggested an alternative

$$e_s = \beta e_{sat}(T_{ss}) + (1-\beta) e_a \qquad (6)$$
$$\text{with} \quad \beta = \tfrac{1}{4}\,[1 - cos\,(\theta/\theta_{fc}\pi)]^2 \qquad \text{if } \theta < \theta_{fc}$$
$$\text{and} \quad \beta = 1 \qquad \text{otherwise.}$$

θ and θ_{fc} are moisture content and field capacity, respectively, of the surface soil layer; and e_a is the vapour pressure of the overlying air. Lee and Pielke found this to be more representative of soil surface conditions, and has the advantage of only requiring one parameter to be estimated (θ_{fc}).

2.4 Transpiration

Transpiration is represented as:

$$\lambda E_{stom} = \rho C_p/\gamma\ (e_{sat}(T_v) - e_b)/(r_v + r_b) \qquad (7)$$

where E_{stom} is transpiration; r_v is the canopy scale stomatal resistance, from here on called "canopy resistance"; and r_b is the aerodynamic resistance between the canopy surface and the canopy air; $e_{sat}(T_v)$ is the saturation vapour pressure at the vegetation temperature, T_v. We use the atmospheric stability scheme of Choudhury [1989]. The dependence of r_v on environmental factors is expressed through its reciprocal, canopy conductance, g_c:

$$g_c = 1/r_v = [f_1(PAR) f_2(D) f_3(T_v) f_4(\psi_l)] \quad (8)$$

The form of these functions will vary with species and whether they are applied to mixed species systems as a whole. Choudhury [1989] expressed these functions as follows:

$$f_1(PAR) = g_{max} + (g_0 - g_{max}) exp(-k_1 \, PAR) \quad (9a)$$

$$f_2(D) = 1 - D/D_{max} \quad (9b)$$

$$f_3(T_v) = 1/[1 + (T_v - T_{opt})/15]^4 \quad (9c)$$

$$f_4(\psi_l) = 1/[1 + (\psi_l / \psi_{1/2})^{\zeta_l}] \quad (9d)$$

where g_o and g_{max} are the minimum and maximum stomatal conductances, respectively; D_{max} is the vapour pressure deficit at which g_c would go to zero; T_{opt} is the optimum temperature at which g_c is a maximum; and (9d) is from Fisher et al. [1981], with $\psi_{1/2}$ the leaf water potential at which g_c is half its maximum, and ζ_l is a parameter controlling the sensitivity of g_c to water status. Choudhury gives a range of literature values for each of these parameters, which are in Table 1. In COUPLE, the effect of soil moisture on g_c is applied as the weighted sum of (9d) for each of the soil moisture stores, with the weighting determined in principle by the root density in each store. More often it may have to be estimated by optimisation.

In a ventilated chamber study enclosing a mature jarrah (*Eucalyptus marginata*) tree, Bartle and Colquhoun [1985] found the dependence of leaf stomatal conductance on D to be an exponential relationship, with the fitted parameter k_D:

$$f_2(D) = exp(-k_D \, D) \quad (10)$$

2.5 Biomass and Canopy Heat Storage Flux

The relatively large volume of air and biomass in the canopy in forests in south-west W.A. means that the heat storage of the canopy air and biomass are important at sub-diurnal time scales. The change in canopy heat storage, J, has three main components: J_a, change of sensible heat storage in the canopy air (including in the moisture in the air); J_λ, latent heat storage flux through change in moisture content of the canopy air; and change heat storage in the canopy biomass (leaves, J_{leaves}, and branches and trunks, J_{trunks}).

$$J = J_a + J_\lambda + J_{leaves} + J_{trunks} \quad (11)$$

Measurements have shown that J_{trunks} was far the largest [Silberstein et al., 2001], and is the only one discussed here. We invoke a new approach by treating the heat storage of the trunk mass in a similar fashion to the soil heat storage, that is, using the "force-restore" approach [Deardorff, 1978], with the tree trunks approximated as a two layered system with one surface exposed. Strictly the cylindrical nature of the system should be accounted for but as a first approximation this is ignored. The equations for the heat balance of the trunks are then:

$$\alpha_t \frac{\partial T_{bark}}{\partial t} = J_{trunk} \frac{2\sqrt{\pi}}{\rho_{bark} C_{bark} \sqrt{\kappa_{bark} \tau_{bark}}} - \frac{2\pi}{\tau_{bark}} (T_{bark} - T_{core}) \quad (12)$$

$$\frac{\partial T_{core}}{\partial t} = J_{trunk} \frac{1}{\rho_{core} C_{core} \sqrt{\kappa_{core} \tau_{core}}} \quad (13)$$

where ρ, C, κ, T, and τ refer to the tree trunk densities, specific heats, thermal diffusivities, temperatures and time constants, respectively; the subscripts ***bark*** and ***core*** refer to the outer and inner layers of the tree trunk, respectively; and $\alpha_t = 1 + \delta_t / d_t$ is to allow for the surface skin effect on the surface of the tree trunks (with skin thickness d_t, and diurnal damping depth d_t), and is analogous to the soil surface skin effect.

3 OBSERVATIONS

COUPLE is a catchment scale representation and does not resolve internal catchment processes. To fully test the model requires measurements of equivalent states and processes at the catchment scale, such as surface temperature and catchment soil moisture storage, and catchment total soil evaporation, transpiration, and runoff. Definition of these variables is complicated with a number of possibilities. For example, Silberstein et al. [1999] tested one estimate of effective surface temperature [Choudhury, 1989] with a very good fit to Landsat-TM data and stream flow data. The focus here is to test, for a natural eucalypt forest, the main process parameterisations taken from the literature.

The natural forest at the site, in south-west W.A., is a dry sclerophyll woodland with two dominant overstorey eucalypt species, and is subject to a Mediterranean climate. Annual rainfall is 1050 mm, 80% of which falls in the months May to October. The observations were taken during a week in October 1993, at the middle of a very wet spring, and in March 1994, after an extremely dry summer [Silberstein et al., 2001].

The model was initialised with parameter values taken from a nearby catchment considered hydrologically similar, and which had a much longer period of streamflow record [Sivapalan et al., 1996]. There are dangers in making the assumption of similarity, but the main concern here is to test the resolution of the surface energy balance. It was considered safe to use these parameters provided the model run was not long enough to allow soil storage errors to affect the results.

The soil surface moisture content and evaporation was measured using microlysimeters, and soil surface vapour pressure was calculated from air humidity and temperature measurements taken 25 cm above the soil surface every hour or so. A full profile core was drilled at the site at the time of each field campaign, giving a measure of total moisture content in the deep (~35 m) unsaturated zone, and giving estimates of θ_{fc} and θ_{sat}.

To calculate hourly transpiration, we used the energy balance residual method, and then subtracted hourly soil evaporation and the vapour component of canopy heat storage [Silberstein et al., 2001]. This is used to calculate g_c:

$$g_c = E_{stom}/\rho D_q \qquad (14)$$

where D_q is the air specific humidity deficit, (or vapour pressure deficit expressed as specific humidity) in the canopy boundary layer. However, without leaf scale measurements, r_v (see Eq 7) cannot be separated from r_b directly to give the dependence of r_v (and g_c) on PAR, D, T_v, and θ independently. To analyse the dependence of r_v+r_b on these variables, (7) is inverted:

$$r_v+r_b = \rho C_p/\gamma\,(e_{sat}(T_v)-e_b)/(\lambda E_{stom}) \qquad (15)$$

Following Choudhury [1989], we estimated that r_b would generally be <10 s m^{-1}, while the minimum sum of r_v+r_b was found to be ~25 s m^{-1} [Silberstein et al, in press], with the majority of values well above this. Canopy resistance of forests is generally much higher (100-500 s m^{-1}, with the ratio of r_v to r_b often being 50 or more [Monteith and Unsworth, 1990]. Hence we neglect the effect of r_b when calculating r_v, and assume the leaves are very close to the ambient canopy air temperature, so $T_v \approx T_b$.

4 RESULTS AND DISCUSSION

4.1 Soil Surface Vapour Pressure

Equation 5 dramatically overestimates the vapour pressure under these soil conditions by nearly an order of magnitude (Figure 1b), while the expression of Lee and Pielke [1992] does very well. In deriving e_s from ψ_s in (5), we used data from Carbon et al., [1980] to derive ψ_o = 0.25 and ξ =1.9. Because θ_{fc}, is somewhat arbitrary, e_s is shown for two values of θ_{fc}, =0.25 (which has r^2 of 0.86) and θ_{fc} =0.35 (r^2=0.93). Under the gradient driven evaporation model (5) results in excessive overestimation of evaporation at the soil surface.

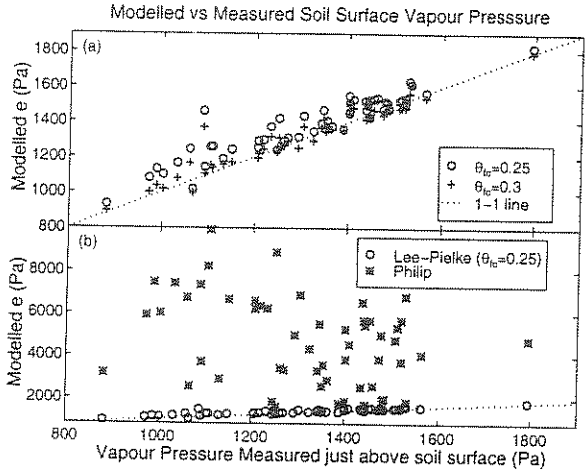

Figure 1. Comparison of the two models for soil surface vapour pressure given by a) Eq (4) and b) Eq (6) with (4) for comparison.

4.2 Canopy Conductance

The relationship between canopy conductance, $g_c=1/r_v$, normalised by LAI, and each environmental variable given in (9a-d) and (10) is shown in Figure 2. The dependence of g_c with each variable independently is represented by the overlying envelope curve; which assumes that at some stage through the observation period conditions will be near optimum for each of the variables. Figure 2 shows that dependence of g_c on VPD is much better described by an exponential decay with increasing VPD, than the linear relationship used by Jarvis [1976] and others [for example, Choudhury, 1989].

PAR is calculated from incoming short-wave radiation [Choudhury, 1989, Eqs-32—34]. Two models [from (9a)] are plotted, one using the best estimates of parameters from those given by Choudhury and one which seemed to envelope the data points best (Table 1), although clearly a majority of points were well under the optimum curve. This forest seems less responsive to radiation levels than those discussed by Choudhury, as the exponent is lower and as the data show few high conductance points at low radiation levels.

The envelope curve for the dependence of g_c on air temperature is better represented by an inverse 2nd power relationship than by the 4th power (9c) given by Choudhury, although we do not have data for temperatures above ~35°C. The optimum temperature (~19°C) is very close to the annual mean temperature for the region. The observed g_c dependence on temperature is complicated by the VPD dependence on temperature. However, when the dependencies on PAR and VPD are compensated for [Silberstein et al., in press], this relationship reverts to being closer to the inverse 4th power with T_{opt}~21°C.

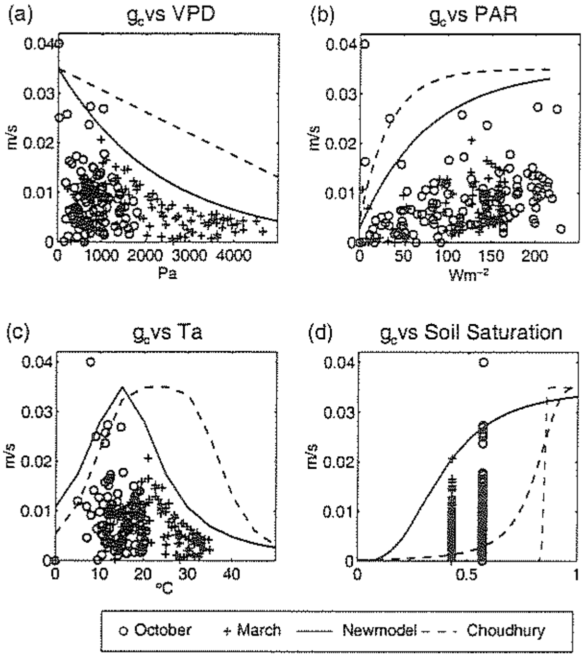

Figure 2. Canopy conductance, g_c, plotted against environmental variables.

Table 1. Canopy conductance parameter values from Choudhury [1989] and from our data.

Function			
$f_1(PAR)$	g_o(ms^{-1})	g_{max}(ms^{-1})	k_1
Our data	0.004	0.036	0.013
Choudhury	0.0008	0.006	0.03
$f_2(VPD)$	D_{max}	D_{min}	k_D(Pa^{-1})
Our data	N/A	N/A	0.00042
Choudhury	7500Pa	2500Pa	N/A
$f_3(T_a)$	T_{opt}		
Our data	20°		
Choudhury	23°		
$f_4(\theta)$	ζ_1	$\Psi_{½}$	
Our data	1	4.5m	
Choudhury	1-15	70-190m	

Because field data were only available for two short periods in two seasons, attempts to generalise the g_c dependence on catchment scale soil moisture are hazardous. The dependence is really only suggested by the data (Figure 2d), where we show all values of g_c for the two periods, and the saturation level of the top 25m of soil. The average soil moisture content in the unsaturated zone (*F* store in the model), taken from the drilled profiles, is used to calculate the bulk soil-root water potential in (5), with ψ_o=0.5 and ξ=3 from data given by Carbon et al. [1980]. The value of $\psi_{½}$ was determined by fitting the curve to pass through the estimated maximum conductance during the two field measurement periods (the extreme outlier being excluded). The nearest two curves from those given by Choudhury are also plotted.

4.3 Latent Heat Flux

When the improved parameterisation is included in the model, the partitioning of latent heat between soil evaporation and transpiration is dramatically improved (Figure 3, for October only). While both versions simulate total transpiration reasonably well, the version with the literature parameterisations dramatically overpredicts soil evaporation, and consequently underpredicts transpiration. The difficulty that (5) (Model A) has in capturing the soil surface vapour pressure, is manifested as a very large overestimation of soil evaporation. This tends to humidify the canopy air and suppress transpiration. The linear dependence of Model A on VPD is very sensitive to the choice of D_{max}. Choudhury [1989] suggests D_{max} should be 7-8 kPa, but this significantly enhances transpiration at low VPD. To match more of the data D_{max} needs to be around 4 kPa, which would result in very restricted summer transpiration when the VPD exceeds this level for considerable lengths of time.

There is a slight but sharp increase in soil evaporation at night, which is due to an artefact of the different time scales of operation between the energy balance and the soil moisture balance. The latter is updated at midnight, and so there is an apparent slight recovery in soil moisture that drives the increase in soil evaporation.

4.4 Canopy Heat Flux

The canopy heat flux is large at the beginning and end of each day (figure not shown for space restrictions). The simple "force-restore" model simulates the trends well, although it is very sensitive to the values set for the trunk diffusivity and, of course, to the estimate of biomass at the site. This is very encouraging because it is simple to implement, and is of a similar magnitude to G.

5 CONCLUSIONS

We have shown that we can simulate energy balance and evaporation fluxes well for a tall and open natural forest woodland. This is the first time this has been done for the forests of south-west Australia. Literature parameterisations are clearly inadequate for this system, and this emphasises the need for care when transposing literature findings to other environments. The Philip [1957] equation for soil surface vapour pressure results in a significant overestimate in soil surface evaporation, whereas that of Lee and Pielke [1992] performed very well. The simple "force-restore" model works well for canopy heat storage, at least as well as for soil heat flux (not discussed here). These findings emphasise the need to obtain appropriate and sufficient data to complement model development.

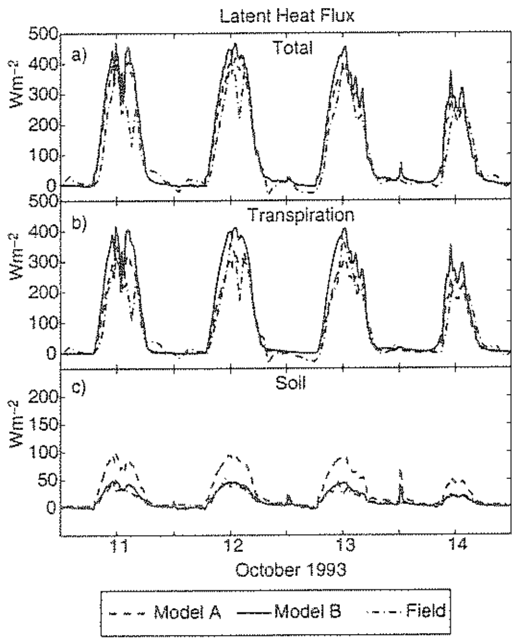

Figure 3. Latent heat flux during 4 days in October, 1993. Model A is the "old" model using literature parameter values, and Model B uses our new parameterisations.

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