An argument for the use of two-layer SVAT schemes to simulate terrestrial carbon dioxide fluxes

C. Huntingford, R.L. Hall and A. Verhoef

Institute of Hydrology, Wallingford, OX10 8BB, U.K.

Abstract

An order-of-magnitude argument shows that the different surface temperatures of soil and sparse vegetation affect carbon assimilation and soil respiration significantly. However, regulation of assimilation through associated modulation of in-canopy carbon dioxide levels is weak.

It is shown that for many vegetation types, the use of a two-layer representation of vegetation is essential to predict, accurately, terrestrial carbon fluxes, primarily through allowing different surface energy balances.

Introduction

Two-layer land-surface models (Shuttleworth and Wallace, 1985; Dolman, 1993; Huntingford *et al.*, 1995) have been used to describe the surface energy balance of a system comprising a plant canopy over either bare soil or an understorey. These models incorporate aerodynamic and surface (bulk stomatal or soil) resistances that differ for each layer and result in different surface temperatures, humidities and local energy partitioning. In reality, these differences can be large: see, for example, the temperature differences in Sahelian savannah measured by Verhoef, (1995). Interaction between the latent and sensible heat fluxes arising from the two layers is simulated within the two-layer framework by combining the aerodynamic resistances from the two layers within the canopy. Such interaction has been demonstrated to be important for the bush-soil combination of Sahelian tiger bush (Blyth and Harding, 1995).

The development of two-layer models to include carbon dioxide fluxes may provide new insights into vegetation processes through the linkages that can be simulated within the two-layer framework. Such linkages may result in different predictions of carbon dioxide fluxes to those from a single-layer model. Recent descriptions of stomatal resistance contain a dependence upon carbon assimilation (Ball *et al.*, 1987, Leuning, 1995) which dependence relates explicitly the surface energy balance to fluxes of carbon dioxide from the vegetation. In addition, the dependence of assimilation rates upon temperature and carbon dioxide concentrations, and of soil respiration upon temperature,

can be simulated within the two-layer network of aerodynamic resistances. The effects of different layer temperatures (themselves a function of the surface energy balances) upon respiration and assimilation can therefore be described, as can the direct effect of respiration upon assimilation through modulation of canopy level carbon dioxide concentration.

The importance of two-layer modelling for describing, accurately, surface energy fluxes has been established. In this paper, feedbacks within the two-layer (vegetation and soil) structure that are associated with the surface carbon dioxide fluxes are isolated and their importance assessed. The use of order-of-magnitude arguments allows quite general conclusions to be drawn.

Carbon assimilation and soil respiration equations

A SIMPLE MODEL OF ASSIMILATION

The model of net assimilation, a_n (mol CO_2 m^{-2}s^{-1}), summarised here, is identical to that of Raupach (1998), which in turn is based upon the model of Farquhar *et al.* (1980).

Assimilation takes the minimum of two equation roots, $a_n = \min(a_T, a_I) - r_d$ where a_T and a_I (both mol CO_2 m^{-2}s^{-1}) are models of photosynthesis and r_d (mol CO_2 m^{-2}s^{-1}) is dark respiration. Both a_T and a_I are limited by leaf temperature, T_s (°C) and by intercellular carbon dioxide concentration c_i (mol CO_2 (mol air)$^{-1}$). Variable c_i satisfies $a_n V_{mol} = (c_s - c_i)/1.6 r_s$ where c_s (mol CO_2 (mol air)$^{-1}$)

is leaf-surface carbon dioxide concentration, r_s (s m^{-1}) is the leaf stomatal resistance for water and V_{mol} (m^3 mol^{-1}) is the molar volume. Variable a_I (sometimes referred to as the light limited solution) is also dependent upon downward shortwave radiation I (W m^{-2}). Leaf stomatal resistance, disregarding soil moisture stress, is given by (Leuning, 1995) $r_s^{-1} = g_0 + ba_n(c_s - 1.1\Gamma^*)^{-1} (1 + D_s/D_{s0})^{-1}$. Based upon typical values given by Leuning (1995) (here in meteorological units), parameters are set as $g_0 = 10^{-3}$ m s^{-1}, b = 0.2 m^3 (mol air)$^{-1}$ and $D_{s0} = 0.01$ kg kg^{-1}. Variable $D_s(T_s, q_s)$ (kg kg^{-1}) is leaf-surface specific humidity deficit where q_s (kg kg^{-1}) is leaf-surface specific humidity. Γ^* (mol CO_2 (mol air)$^{-1}$) is the temperature dependent CO_2 compensation point.

Elimination of min(a_T, a_I), c_i and g_s gives a single equation relating net leaf assimilation to environmental variables; $a_n = a_n(T_s, c_s, D_s(T_s, q_s), I)$. Canopy assimilation is assumed to be linear in leaf area index, L (m^2 m^{-2}), i.e. given by a_nL, which is valid for sparse vegetation (L approximately less than two). The dependences of a_nL upon leaf temperature and leaf-surface carbon dioxide concentration (for single representative values of leaf area index, humidity and different values of downward shortwave radiation) are presented in Fig. 1a and Fig. 1b respectively.

SOIL RESPIRATION

Soil respiration, R_{soil}, (mol CO_2 m^{-2}s^{-1}) is dependent upon soil temperature (see e.g. Lloyd and Taylor, 1994) and is often assumed to be dependent upon leaf area index (as a proxy for root density) and soil moisture, or both (Norman et al., 1992). Here, R_{soil} is given by $R_{soil} = aL \exp\{T_{soil}/T_0\}$ which was used by Verhoef and Allen (1998) (assuming small variability of soil moisture) for Sahelian savannah, and where T_{soil} (°C) is the soil surface temperature. On the basis of a range of soil respiration rates quoted by Verhoef et al. (1996), parameter values are assigned as $a = 6.0 \times 10^{-7}$ (mol CO_2 m^{-2}s^{-1}) and $T_0 = 20.0$ °C. The temperature response of this model of soil respiration is presented in Fig. 1c.

Order-of-magnitude analysis

The aerodynamic resistance networks for sensible heat and carbon dioxide fluxes are shown in Fig. 2. Resistance r_a (s m^{-1}) is between a mid-canopy mixing point (see Shuttleworth and Wallace, 1985) and a reference height above the canopy (an automatic weather station or first model level in a numerical experiment). r_{bs} (s m^{-1}) and r_{bsoil} (s m^{-1}) are bulk boundary-layer resistances (to heat) close to the vegetation and soil layers respectively and r_{au} (s m^{-1}) is an in-canopy resistance. The two-layer model structure, with leaf stomatal resistance as given above, implicitly couples all three fluxes, vapour, heat and carbon dioxide, with each other, and with the associated surface scalar values of

humidity, temperature and carbon dioxide concentration.

The full coupling makes it difficult to assess internal feedbacks, as no single quantity is a simple model diagnostic. However, order-of-magnitude arguments provide a technique for assessing the strength of selected feedbacks in isolation, a process that cannot be achieved by analysis of the full two-layer model. A physically realistic value is assigned to a particular system variable of interest. The governing equations are solved sequentially and a new value obtained for that variable. The difference between the original and newly calculated values of the variable indicates the importance of internal model feedbacks upon that quantity. This analysis technique can be quite general with deductions that are independent of exact specification of site specific model parameters.

This algorithm is now applied to the two-layer model equations. The vegetation structure selected for analysis follows the model of Shuttleworth and Wallace (1985), that is a canopy completely covering bare soil. Algebraic manipulation shows that the surface temperature difference, ΔT (°C) between the two layers is given by:

$$\Delta T = T_{soil} - T_s = \frac{(r_{au} + r_{bsoil})H_2 - r_{bs}H_1}{\rho c_p} \quad (1)$$

where H_1 (W m^{-2}) is the sensible heat flux from the upper canopy, H_2 (W m^{-2}) is the sensible heat flux from the soil, ρ (kg m^{-3}) is air density and c_p (J kg^{-1} K^{-1}) is the specific heat capacity of air. Similarly, the difference in carbon dioxide concentration, Δc (mol CO_2 (mol air)$^{-1}$), satisfies

$$\begin{aligned} \Delta c &= c_{soil} - c_s \\ &= R_{soil}V_{mol}(r_{au} + 1.4r_{bsoil}) + 1.4a_nLV_{mol}r_{bs} \end{aligned} \quad (2)$$

where c_{soil} (mol CO_2 (mol air)$^{-1}$) is the carbon dioxide concentration at soil level. If the vegetation and soil are treated as a single layer, then predicted surface temperature and carbon dioxide levels will lie between the soil and canopy two-layer values. Therefore, the magnitudes of ΔT and Δc are representative of predicted changes in surface values of temperature and carbon dioxide concentration that result from replacing a single-layer model by a two-layer description.

From a priori knowledge of typical surface energy fluxes, $H_1 \sim O(10^2)$, $H_2 \sim O(10^2)$ and Fig. 1a–c shows that both $a_nL \sim O(10^{-5})$ and $R_{soil} \sim O(10^{-5})$ (where '$\sim O$' means 'of the order of'). It is also known that $\rho \sim O(1)$, $c_p \sim O(10^3)$, $V_m \sim O(10^{-2})$, $L \sim O(1)$ and so from the dominant terms of Eqns. (1) and (2), $\Delta T \sim O(10^{-1} \times \max[r_{au} + r_{bsoil}, r_{bs}])$ and $\Delta c \sim O(10^{-7} \times \max[r_{au} + 1.4r_{bsoil}, r_{bs}])$. The impact that a temperature difference of magnitude ΔT has upon canopy assimilation a_nL and soil respiration R_{soil} are expressed in percentage terms by statistics $\lambda(a_nL)_{\Delta T}$ and $\lambda(R_{soil})_{\Delta T}$ respectively. Similarly, the effect of a change in surface carbon dioxide concentration of magnitude Δc upon assimilation is expressed as $\lambda(a_nL)_{\Delta c}$. These statistics
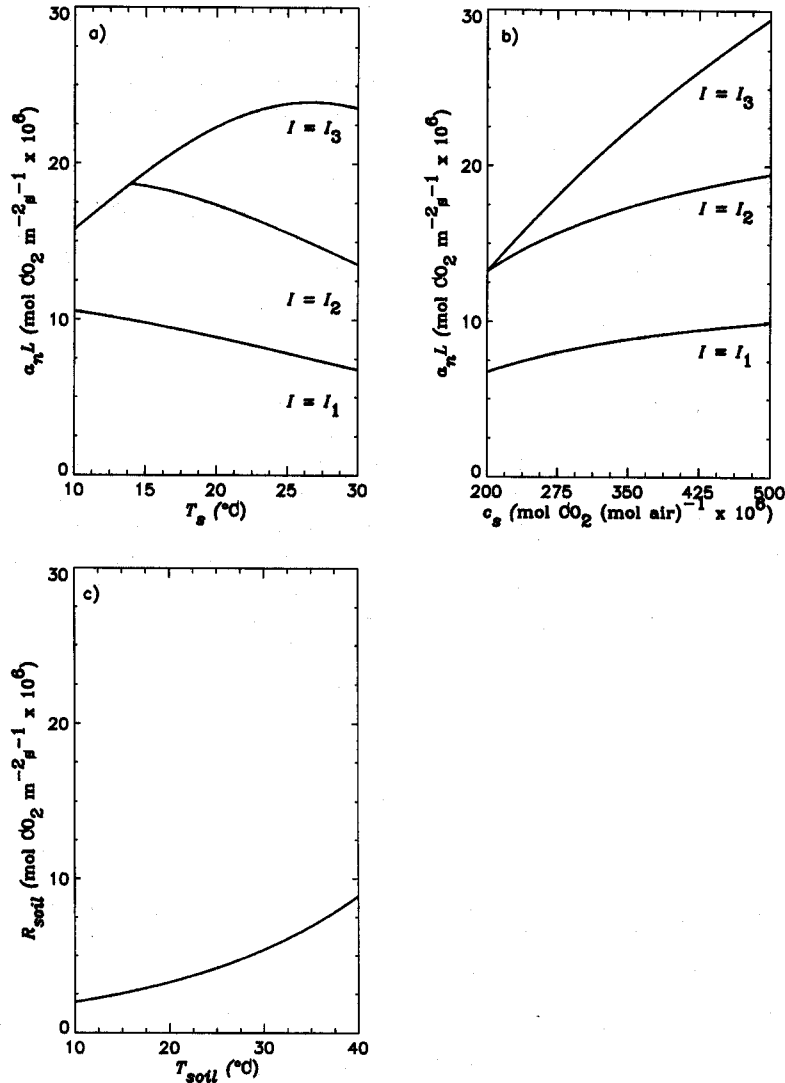
Fig. 1. *The response of net assimilation, a_nL to: (a) vegetation surface temperature, T_s: (b) vegetation surface carbon dioxide level, c_s; and (c) the response of soil respiration, R_{soil} to soil surface temperature, T_{soil}. Values of T_{soil} cover a higher range than T_s to reflect the expected higher soil temperatures. Control driving variable values are $q_s = 6 \times 10^{-3}$ kg kg^{-1} and $L = 2$. Further, in Fig. 1(a), $c_s = 350 \times 10^{-6}$ mol CO_2 (mol air)$^{-1}$ and in Fig. 1(b), $T_s = 20$ °C. Incoming shortwave radiation is $I_1 = 200$ W m^{-2}, $I_2 = 400$ W m^{-2} and $I_3 = 1000$ W m^{-2}. Solution $I = I_1$ is always light limited, $I = I_3$ is always temperature limited and $I = I_2$ is temperature limited at lower temperatures and light limited otherwise.*

are defined as the magnitude of expected change upon a_nL and R_{soil} divided by representative values. That is:

$$\lambda(a_nL)_{\Delta T} = 100 \frac{\Delta T \dfrac{d(a_nL)}{dT_s}}{a_nL},$$

$$\lambda(R_{soil})_{\Delta T} = 100 \frac{\Delta T \dfrac{dR_{soil}}{dT_{soil}}}{R_{soil}},$$

$$\lambda(a_nL)_{\Delta c} = 100 \frac{\Delta c \dfrac{d(a_nL)}{dc_s}}{a_nL}.$$

Figure 1 shows $a_nL \sim O(10^{-5})$, $R_{soil} \sim O(10^{-5})$, and gradients may be as large as $d(a_nL)/dT_s \sim O(10^{-6})$, $dR_{soil}/dT_{soil} \sim O(10^{-6})$, $d(a_nL)/dc_s \sim O(10^{-1})$.

It remains to estimate the magnitudes of the aerodynamic resistances. Shuttleworth and Wallace (1985) calculate a value as high as $r_{au} = 128$ sm^{-1} for a typical agricultural crop (their model does not include r_{bsoil}). Leaf boundary-layer resistance and bulk soil boundary-layer resistance are of the form $A\surd(d/u)$ where A (s$^{1/2}$ m^{-1}) is an empirical constant of $O(100)$ (Huntingford et al., 1995, McNaughton and Van Den Hurk, 1995), d (m) is a characteristic length scale and u (ms^{-1}) a local windpseed. For leaves of width $d \sim O(10^{-2})$ and a windspeed of $u \sim O(1)$, then $r_{bs} \sim O(10) \times L^{-1}$ where L^{-1} scales from leaf to bulk boundary-layer resistance, i.e. $r_{bs} \sim O(10)$. For soil, with clod size as the characteristic length (Goudriaan, 1977), $d \sim O(10^{-2})$ and $u \sim O(10^{-1})$, then r_{bsoil} is between $O(10)$ and $O(100)$. Hence $\max[r_{au} + r_{bsoil}, r_{bs}] \sim O(100)$ (the factor 1.4 taken as $O(1)$ for CO_2 fluxes), then $\Delta T \sim O(10)$ and
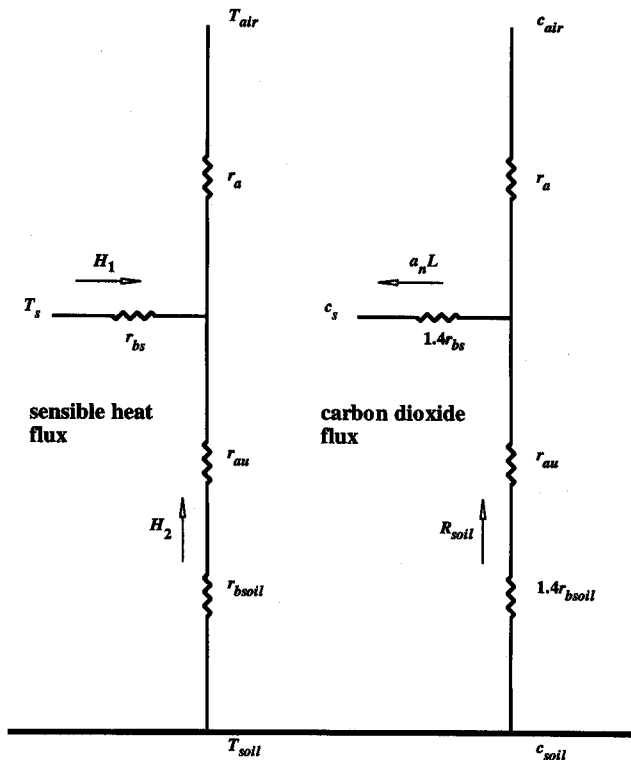
Fig. 2. *A schematic of the aerodynamic resistances involved in two-layer modelling of heat and carbon dioxide fluxes. The factor of 1.4 arises from the decreased diffusivity of carbon dioxide compared to heat across laminar boundary-layers adjacent to surfaces. Variables T_{air} (°C) and c_{air} (mol CO_2 (mol air)$^{-1}$) are respectively values of temperature and carbon dioxide concentration at the reference level.*

$\Delta c \sim O(10^{-5})$. This gives $\lambda(a_nL)_{\Delta T} \sim O(100\%)$, $\lambda(R_{bsoil})_{\Delta T} \sim O(100\%)$ and $\lambda(a_nL)_{\Delta c} \sim O(10\%)$. So with all other factors constant, modelling different surface temperatures with a two-layer model may adjust predictions of both assimilation and soil respiration by as much as their original single-layer model values, although feedbacks directly through canopy carbon dioxide levels are smaller.

Conclusions

A two-layer model of vegetation explicitly allows two surfaces to maintain different energy balances. An order-of-magnitude argument reveals that if a physically representative two-layer structure is used to model carbon dioxide fluxes, then predictions of both assimilation and soil respiration may differ significantly from those of a single-layer model. This behaviour is attributed primarily to large differences between canopy and soil temperatures which can be simulated within a two-layer model; the influence of carbon dioxide fluxes within the canopy upon photosynthesis is small. It is recognised that large temperature differences will not occur for all vegetation systems and microclimates, but the parameter values and conditions chosen within this paper are not unrealistic.

For many vegetation types, other authors have demonstrated, the necessity to use two-layer models to replicate, accurately, sensible and latent heat fluxes. This paper demonstrates that this is also true for terrestrial carbon dioxide fluxes.

Acknowledgements

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