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Energy and water flow through the soil–vegetation–atmosphere system: the fiction of measurements and the reality of models

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Abstract

This paper summarizes basic concepts and definitions in models of the SVA system and then emphasizes inconsistencies between model variables and observations for the soil, vegetation and atmosphere elements. This is done first in a qualitative sense, then analytically for the observations of the radiometric temperature of vegetation canopies.

Notwithstanding the significant efforts dedicated to systematic comparison of models and to performing complex experiments to construct a rich data base for model validation, the inter-model variability of predicted fluxes and state variables remains large and it is hard to pin-point specific causes. The argument developed in this paper is that a different avenue should be explored in search of a solution, namely an in-depth analysis of the nature of feasible observations of the SVA system, in order to detect and understand inconsistencies in model variables and parameterizations. Model equations define state variables and parameters rather precisely, while parameterizations are often established from experimental data, assuming that observed and model variables are consistent. The latter cannot be taken for granted, however. There are now tools of investigation that were not available in the early years of land-surface science. We are now able to construct detailed and realistic 3D models of elements of the SVA system, e.g. a soil-foliage system, and to model radiative and convective processes in such 3D systems. The latter gives us the opportunity of modeling observations of elements of the SVA system, such as the soil matrix or a vegetation canopy, in a rather realistic way. When dealing with highly heterogeneous systems, this capability provides ways and means to understand how the integral magnitudes we measure, such as soil electrical resistivity or radiance emitted by a canopy, relate to the object properties we seek to determine.

Introduction

The argument developed in this paper is that careful consideration of the nature of observations would lead to better models of the Soil–Vegetation–Atmosphere (SVA) system. Such models have developed from concepts rather than experimental physics,

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and this led to a significant gap between the nature of variables in models and variables accessible to measurements. Some definitions and concepts are summarized briefly in this Introduction.

Land Surface Models (LSMs) describe physiological and biophysical processes as well as soil biochemical and physical processes. Exchanges with the atmosphere are described by a Soil–Vegetation–Atmosphere Transfer (SVAT) model. In these models, state variables such as temperature, moisture and nitrogen content vary with time, but all other properties of the system are supposed to be stable (e.g. Brisson et al. 1998; Spitters, Van Keulen and Van Kraalingen 1989; Hoogenboom, Dekker and Althuis 1998). Some of these LSMs also describe radiative transfer in the Soil–Vegetation–Atmosphere system (e.g. Weiss et al. 2001; Schneider 1999).

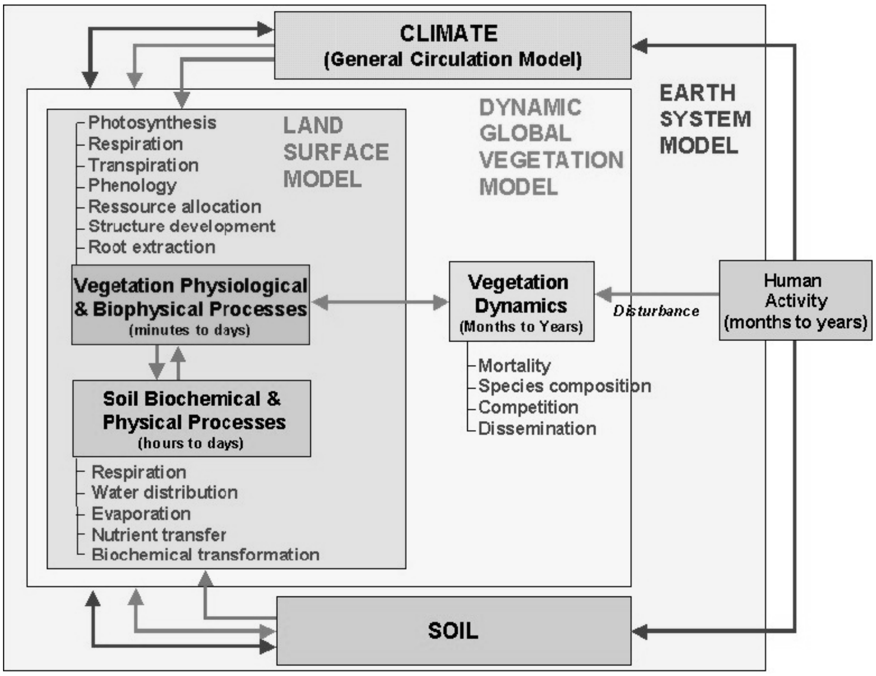

Figure 1. Schematic of the processes determining the interactions between vegetation and the climate system, at various space and time scales and of the types of models needed to describe them

An additional challenge is spatial scale: models describing the interactions of the atmosphere with the terrestrial biosphere at the global scale use LSMs at spatial scales of 50 km or larger assuming that both processes and variables are scale-invariant. The latter implies that the relation of observations with model variables is assumed to be the same at all spatial scales.

Dynamic Global Vegetation Models (DGVMs) typically include an LSM and an additional module that describes the dynamic evolution of the ecosystem in response to climate. Interactions with soil properties may be included to describe the dynamics of variables other than temperature, moisture and nitrogen content. They differ significantly in complexity and reliance on parameterizations. Simpler models are more efficient, but also more limited to describe particular processes. They are suitable for long-term studies or to investigate basic feedbacks in coupled Earth System Models (Cox et al. 2000). On the other hand, more complex models allow for direct studies of ecophysiological processes and their implications on a global scale. Cramer et al. (2001) recently reviewed and compared various DGVMs. The differences in model complexity and reliance on parameterization lead to a significant inter-model variability in predicted evolution.

Earth System Models (ESMs) attempt to couple all the relevant processes together, with an emphasis on the atmosphere–surface interactions. They focus on the global spatial scale and are used to simulate the evolution of the entire system over a wide range of time scales. These models necessarily rely on effective parameterizations of small-scale processes.

The issue of parameterization of SVA processes is normally seen in relation with the accuracy of the models where they are applied. The point of view taken here is that a more fundamental issue is whether variables in models and parameterizations are consistent, in the sense of experimental physics, with the variables accessible to feasible measurements. The latter has particular implications when considering remote sensing of the land surface as a source of observations in this context.

The exploitation of remote-sensing observations in the context of modeling earth-system processes can follow two different approaches. In the first one, *forcing*, biophysical variables (e.g., fAPAR, LAI, soil temperature, etc.) are determined from radiance measurements and included in the input data stream. The second approach consists of the *assimilation* of either biophysical variables or radiometric measurements into dynamic models. This is achieved by adjusting one or more model parameters until the model matches the observations. In both cases it is assumed that model and observed variables are identical. The latter is far from self-evident, as argued in this paper.

How global models describe heterogeneous terrestrial vegetation

Spatial heterogeneity is a defining feature that distinguishes terrestrial surfaces from the oceans and the atmosphere. Reliable and accurate fluxes of carbon, water and energy are difficult to estimate at the regional to global scale because of the heterogeneity of the landscapes.

Recent literature demonstrates that forecasts of the long-term evolution of climate depend quantitatively and qualitatively on how the response of terrestrial vegetation is parameterized. For example, Cox et al. (2000) demonstrated that plausible assumptions on the ratio of photosynthesis to respiration lead to substantially different evolutions of the global climate in response to increased CO_2 concentration in the atmosphere. Similarly, Claussen (1997) showed that different representations of biosphere processes may or may not lead to re-growth of vegetation in the Sahara in response to long-term climate trends.

From the point of view of modeling, we may distinguish two main approaches to represent land–atmosphere interactions in global models:

- ‘Frozen biosphere’: A map of global biomes is used to determine the abundance of each land-cover type for all model grid boxes. The exchanges of energy, water and carbon within each grid box are computed using a Land Surface Model (LSM; e.g. Figure 2a). The properties of each biome are time-independent and are specified, typically in tabular form.
- ‘Interactive biosphere’: The LSMs in this category are far more complex than in the previous category. Some of them only describe photosynthesis and respiration, while others characterize the state of the biomes in great detail, including species composition and its evolution in response to climate forcing.

The LSMs implemented in global Earth System Models relate to entire regions, rather than samples of truly homogeneous biomes characterized by well-defined, observable biosphere properties. While all relevant processes may be included in a model of the type depicted in Figures 1 and 2a, such models lump the underlying heterogeneity and non-linearity of terrestrial biosphere processes into variables and parameterizations at the scale of a model grid, say 50 km or larger.

Processes in a complex vegetation canopy may be represented in a fully conceptual way as in Figure 2a. Relevant processes are those determining the interaction of the terrestrial biosphere with the atmosphere, and they are described by simplified equations (parameterizations). Variables appearing in such parameterizations relate to actual canopy properties and state variables, but abstract from the complexity of plant (Figure 2b) and canopy architecture (Figure 2c) and the significant heterogeneity of energy and water fluxes within the canopy space, i.e. at spatial scales significantly smaller than the spatial scale at which the LSMs are used in Earth System Models. Terrestrial biosphere processes may be described by a different class of models, in which processes in the plant environment are described taking canopy architecture into account more precisely (Bouman et al. 1996).

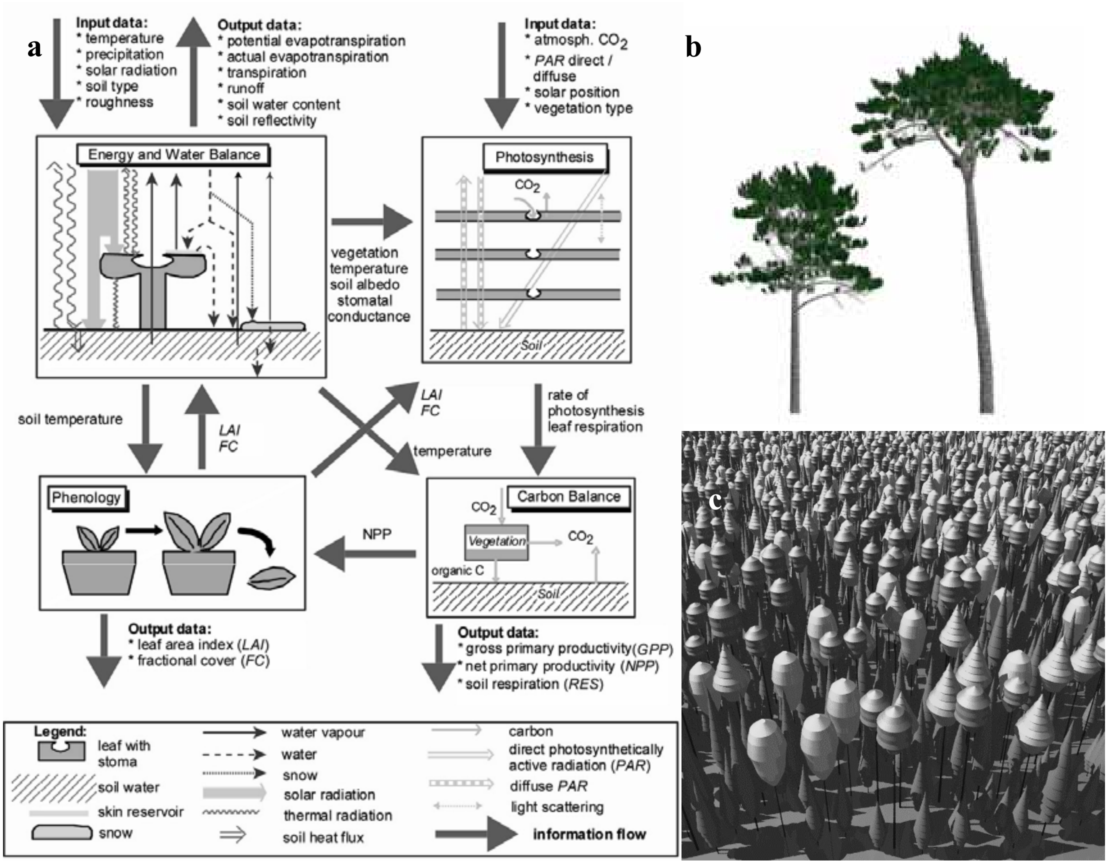

Figure 2. Components of the soil–vegetation–atmosphere system and their interactions: a) schematic of fluxes and variables (Knorr and Heimann 2001); b) computer-generated vegetation canopy (courtesy J.Helbert, Noveltis); c) computer-generated forest (courtesy J.L. Widlowski, IES)

In other words, variables in LSMs, DGVMs and ESMs are assumed constant over length scales much larger than the inherent length scales (see. e.g., Figure 2b and c) of the landscape elements and processes they are meant to describe. Variables and processes are defined in a rather precise way by model equations, but the latter definitions may or may not lead to observable variables. On the other hand observations of the terrestrial biosphere do capture its complex structure. The latter applies also to radiometric observations from space- and airborne platforms, due to the combination of spatial resolution and observation geometry (Line Of Sight, LOS, and Line Of Illumination, LOI). The consequence is that observed variables are inherently different from variables defined by model equations. The latter provide a self-contained and well understood description of processes in the SVA continuum, however, while in most instances we have an intuitive and poor understanding of the precise nature of measurements.

This paper elaborates this concept in relation with the three elements, i.e. soil, canopy space and convective boundary layer, of the system by focusing on the following measurements and processes:

- soil pore space in relation to soil hydrologic properties and soil water flow;
- canopy space in relation to surface reflectance, temperature, energy and water fluxes at the land – atmosphere interface;
- convective boundary layer in relation to air temperature, humidity, energy and water transport.

Background and theory

Flow of water and energy through the SVA continuum is driven by radiant energy and is described (see, e.g., Figure 2a) by neglecting the actual distribution of plant organs in the soil and in the canopy space. The latter implies neglecting the actual distributions of sources and sinks of water and energy. Water extraction from soil water storage is typically taken into account by means of conceptual sub-models or parameterizations. Changes in heat storage are typically neglected. In other words, model equations are developed and combined by analysing mechanisms in water and energy transfer, and selecting relevant processes. This approach leads to definitions of SVA variables that may or may not be observable. For example, a broad class of LSMs describe exchanges of water, energy and carbon between vegetation and atmosphere by postulating that foliage in a 3D canopy may be replaced by a single leaf.

Water and energy (and CO_2) transfer within a representative volume of the SVA system must comply with the conservation equation for water and heat (and CO_2). For all three scalars the equation reads:

$$\frac{\partial C}{\partial t}+U\frac{\partial C}{\partial x}+V\frac{\partial C}{\partial y}+\frac{\partial F}{\partial z}=S \tag{1}$$

where C is the mean scalar concentration field, F is the vertical scalar flux density, U and V are the wind speed along x and y directions, respectively, with $u^2=U^2+V^2$, u is the horizontal mean wind speed. The coordinates x and z lie in the mean stream-wise and vertical directions, respectively. S is the source density of the concerned scalar. For unsteady conditions in an extensive, horizontally homogeneous canopy in which horizontal variations in gradients of air temperature and water vapor are ignored, Eq.1 is dominated by the vertical flux divergence and source terms, and reduces to:

$$\frac{\partial C}{\partial t}+\frac{\partial F}{\partial z}=S \tag{2}$$

Such an equation applies to heat and water movement from below the root zone to the atmosphere above the canopy by substituting appropriate scalar variables and the associated capacitances, conductivities and sources. The source or sink distributions are derived by the local heat and water balance in each point within the representative volume of the SVA. In the three elements of the soil–vegetation–atmosphere system the preceding equations take slightly different forms as indicated below. To illustrate the assumptions underlying the conservation equations Eqs. 1 and 2, the forms of these equations applying to heat and water transfer in the soil, canopy and atmosphere are recalled and briefly discussed below.

Heat and water transfer in the soil

The conservation equation for heat and water content in the soil (1D case, see Eq. 2) is expressed as (Norman and Campbell 1983):

$$c_{\mathrm{s}}\frac{\partial T_{\mathrm{s}}(\mathrm{z})}{\partial t}=\frac{\partial}{\partial z}\left(K_{\mathrm{D}}(z)\frac{\partial T_{\mathrm{s}}(\mathrm{z})}{\partial z}\right)+S_{\mathrm{SH}}(\mathrm{z}) \tag{3}$$

$$\frac{\partial W(z)}{\partial t}=\frac{\partial}{\partial z}\left(K_{\mathrm{w}}(\mathrm{z})\frac{\partial \psi}{\partial z}-k_{\mathrm{w}}g\right)+S_{\mathrm{SW}}(\mathrm{z}) \tag{4}$$

where c_{s} is volumetric heat capacity of soil (J m^{-3} K^{-1}), $K_{\mathrm{D}}(z)$ is soil thermal conductivity (Wm^{-1}K^{-1}), $S_{\mathrm{SH}}(z)$ is the source/sink for heat including phase transitions of water (liquid – vapor and ice – liquid, W is the volumetric water content, $K_{\mathrm{w}}(z)$ is the capillary conductivity (kg s m^{-3}), ψ is the soil metric potential (J kg^{-1}), $S_{\mathrm{SW}}(z)$ is a source/sink including root uptake, thermally induced vapor flow or, in the surface layer, the difference between infiltration and soil evaporation.

In the soil, $K_{\mathrm{D}}(z)$ is a function of soil type and water content, and c_{s} is mainly a function of water content. Norman and Campbell (1983) have given a detailed description about the water movement and the associated parameters in the soil. Besides conservation of (water) mass and energy, Eqs. 3 and 4 express the principle that water flows from wetter to drier locations and heat from warmer to colder locations.

Heat and water-vapor transfer in the canopy space

The conservation equations for heat and water content in the space occupied by a vegetation canopy (1D case, see Eq. 2) read:

$$\rho_{\mathrm{a}}c_{\mathrm{p}}\frac{\partial T_{\mathrm{ac}}(z)}{\partial t}=\frac{\partial}{\partial z}\left(K_{\mathrm{h}}(z)\frac{\partial T_{\mathrm{ac}}(z)}{\partial z}\right)+S_{\mathrm{H}}(\vec{\mathbf{r}}) \tag{5}$$

$$\frac{\rho_{\mathrm{a}}\varepsilon}{p}\frac{\partial e_{\mathrm{ac}}(z)}{\partial t}=\frac{\partial}{\partial z}\left(\frac{\varepsilon K_{\mathrm{h}}(z)}{c_{\mathrm{p}}p}\frac{\partial e_{\mathrm{ac}}(z)}{\partial z}\right)+S_{\mathrm{E}}(\vec{\mathbf{r}}) \tag{6}$$

where $\varepsilon K_{\mathrm{h}}(z)/(c_p p)$ is the eddy diffusivity for water vapor (kg m^{-1} s^{-1} Pa^{-1}), $S_{\mathrm{H}}(\vec{\mathbf{r}})$ is the source/sink distribution of heat(W m^{-3}), and $S_{\mathrm{E}}(\vec{\mathbf{r}})$ is the source/sink distribution for water vapor (kg m^{-3} s^{-1}). $S_{\mathrm{H}}(\vec{\mathbf{r}})$ and $S_{\mathrm{E}}(\vec{\mathbf{r}})$ are determined by the energy balance of leaves at each point (located at $\vec{\mathbf{r}}$).

Heat and water transfer in the atmosphere above the canopy surface

In the atmospheric surface layer, the conservation equations for heat and water vapor (1D case, see Eq. 2) are:

$$\rho_a c_p \frac{\partial T_a(z)}{\partial t} = \frac{\partial}{\partial z}\left(K_h(z) \frac{\partial T_a(z)}{\partial z} \right) \tag{7}$$

$$\frac{\rho_a \varepsilon}{p} \frac{\partial e_a(z)}{\partial t} = \frac{\partial}{\partial z}\left(\frac{\varepsilon K_h(z)}{c_p p} \frac{\partial e_a(z)}{\partial z} \right) \tag{8}$$

where $K_h(z)$ is:

$$K_h(z) = \frac{k u_*(z-d)}{\Phi_h\left(\frac{z-d}{L}\right)} \tag{9}$$

the friction velocity u_* (m s^{-1}) is:

$$u_* = \frac{k u(z)}{ln\left(\frac{z-d}{z_{0m}}\right) - \Psi_m\left(\frac{z-d}{L}\right)} \tag{10}$$

and L is the Monin-Obukhov length, defined as:

$$L = -\frac{\rho_a c_p u_*^3 \theta_v}{kgH} \tag{11}$$

Actual structure of soil, canopy space and atmospheric surface layer

For all the three elements of the SVA system, variables and coefficients of the conservation equations 3 through 8 are defined over a continuous domain. On the other hand, real-world soil, vegetation canopy and atmospheric surface layer are inherently 3D. A complex conceptual system of definitions has been developed over the last 30 years to deal with this apparently unsolvable problem. This system relies heavily on two generic assumptions:

A. It is always possible to define a Minimum Representative Volume (MRV) of soil, canopy and atmospheric surface layer such that Eqs. 3 – 8 apply to variables and parameters averaged over the MRV;
B. The variables and parameters averaged over the MRV can be observed with available instruments.

To verify assumptions A and B in general or to understand under which conditions they may be correct requires a very precise and detailed characterization of the geometrical structure and of local (i.e. at the true length scale of variability) properties of soil vegetation and atmospheric surface layer. To some extent this may be feasible, at the price of rather significant experimental effort, as shown below using a few examples.

Soil

The state of the soil is described in Eqs. 3 and 4 by three continuous variables: T_s, W and ψ, with the latter dependent on W. Flow of heat and water is determined by complex processes and interactions in the pore space (see Figure 3), which depend on the shape, size and mutual position of both pores and grains. Spatial arrangement of pores and grains determines both heat and water flow since: a) the connectivity of the pore space determines the permeability, and b) the extent and degree of contact of grains determines the thermal conductivity. In this respect the mutual arrangement of larger pores and larger grains is important. In clay soils the interaction between the surface of particles and water is an important determinant of soil water flow. Macroscopic (i.e. at the scale of the MRV) state variables, i.e. T_s, W and ψ, and properties, i.e. K_D and K_W, are assumed to account for processes and interactions at the scale of pores. It should be noted that this assumption is manifold:

(a) heat and water flow processes at the pore and grain scale may be neglected and replaced by flux–gradient relationships at the scale of the MRV;

(b) both thermal and capillary conductivity can be measured at the scale of the MRV in a way consistent with (a);

c) state variables at the pore and grain scale may be replaced by state variables at the scale of the MRV.

The complexity of the spatial arrangement of pores and grains suggests that the three aspects (a) through (c) should be evaluated by means of measurements and numerical experiments.

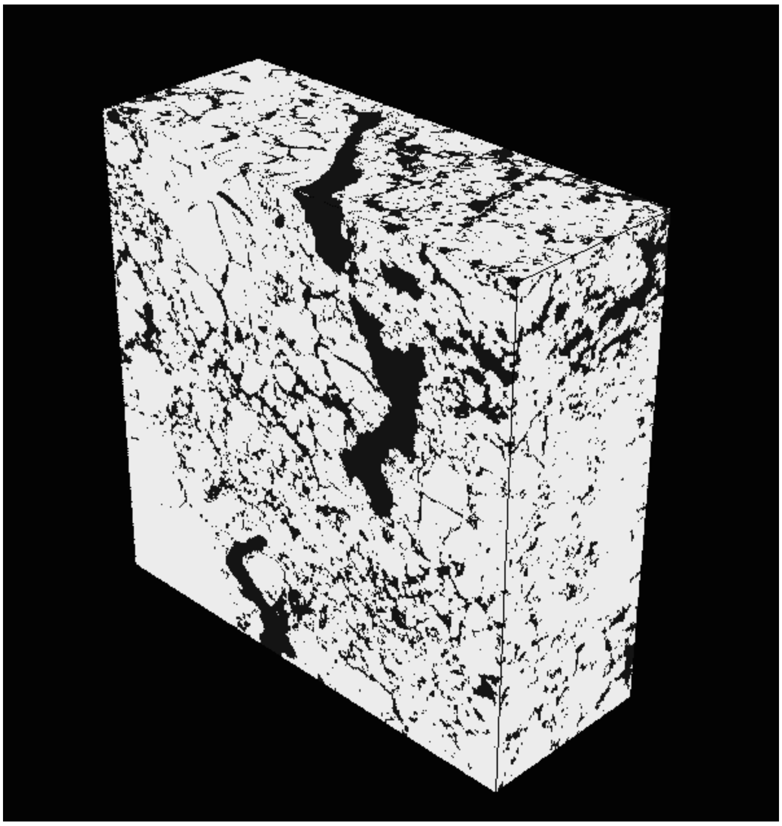

Figure 3. Micro-tomography of a soil sample obtained by imaging a sequence of thin sections; sample size is 2 cm x 5 cm x 5 cm; spatial resolution is 3μ in the x, y and z directions; pore space is black (courtesy of G. Mele, CNR)

Vegetation canopy

Two variables in Eqs. 5 and 6 describe the state of the canopy space: T_{ac} and e_{ac}. The strength of heat sources and sinks depends on the temperature of leaves at any point (see Jia in press), while the difference in e_{ac} between air in the stomata and the canopy space determines transpiration and condensation. Experimental observation of system properties, e.g., leaf boundary-layer resistance is challenging if at all feasible (see e.g. Stanghellini 1987). The temperature of leaves and soil in the canopy space may be measured with, e.g., Thermal InfraRed (TIR) radiometers. The significant thermal heterogeneity can be documented using a TIR camera (Figure 4, see Color pages elsewhere in this book). These observations may be used to assess whether assumptions A and B hold, at least as regards the term S_H in Eq. 5. The latter relates foliage – air heat exchange to the temperature difference between the air in the canopy space and the leaves as:

$$S_H(\vec{r}) = c_H(d, v_a, T_a, T_f) \sum_{i=1}^{N(r)} (T_a - T_f^i) \tag{12}$$

where c_H is a heat-transfer coefficient depending on: leaf size, d; wind speed in the canopy space, v_a; air temperature, T_a; and foliage temperature, T_f. N(r) is the number of leaves at point *r* in the canopy space. This parameterization of foliage – air heat exchange is consistent with the nature of observations: the temperature of individual leaves, or even of portions of a leaf, can be determined with image data such as the ones shown in Figure 4 (see Color pages elsewhere in this book). On the other hand, the determination of leaf size and wind speed in the canopy space is challenging, particularly for heterogeneous canopies. Simpler models of heat and vapor transfer in the canopy space have been developed (see following section) at the price of assumptions on the degree of heterogeneity of foliage temperature.

Atmospheric surface layer

To describe transfer of energy, water and CO_2 in the atmospheric surface layer with Eqs. 7 and 8 it must be assumed that atmospheric properties change only with height. This requires neglecting fast changes in air temperature and humidity, and therefore in all scalar fluxes, due to turbulence. Such changes occur at rather small spatial scales, i.e. spatial variability of water vapor at a given time is rather significant (Figure 5). A more realistic description of the structure and dynamics of the atmospheric surface layer is obtained by Large Eddy Simulation (LES). The LES model of Cuijpers and Duynkerke (1993) was used (Siebersma 2002) to compute detailed 3D fields of temperature, and water vapor in the Convective Boundary Layer (CBL) at spatial resolution of 25 m in the x and y directions and 40 m in the vertical direction. These synthetic data have a twofold relevance: a) true temporal and spatial scales of variability in the CBL and of terrestrial biomes are comparable; b) the conservation equations (Eqs. 7 and 8) apply to a conceptual model of the CBL that requires measurements of average CBL properties, if it would be possible to measure air humidity at the same x,y,z-spatial resolution as the synthetic data in Figure 5.

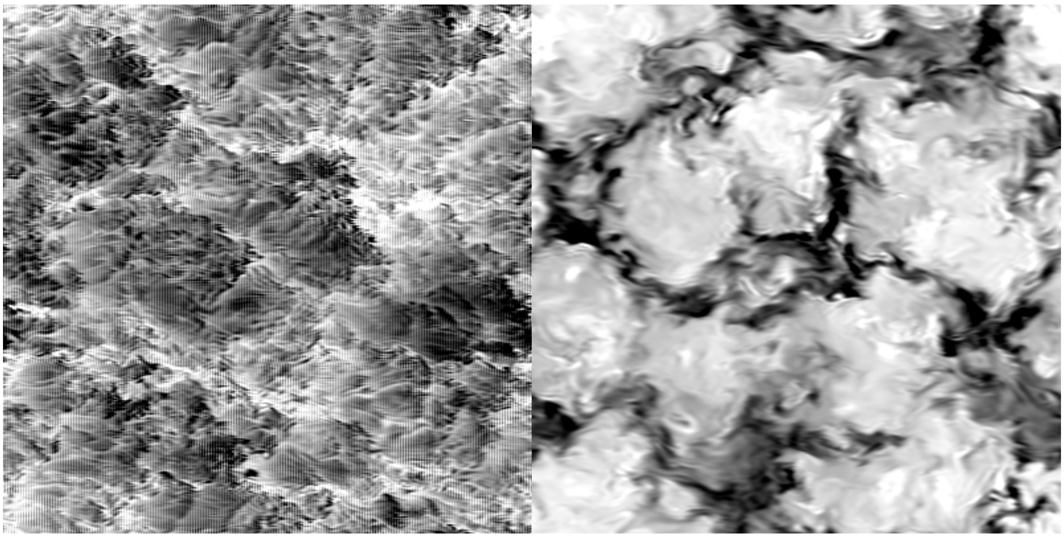

Figure 5. Large Eddy Simulation of water-vapor concentration in the convective boundary layer over a domain of 10 km x 10 km at a horizontal spatial resolution of 25 m: a) surface; b) 3200 m (courtesy of Siebersma, KNMI)

One general feature of patterns (see Figures 3, 4 and 5) in soil, canopy space and surface layer is that they seem to have an identifiable length scale (which may not be easily measured) and they are replicated at larger lengths. The latter implies that: a) the MRV is determined by the inherent length scale and b) spatial variability, particularly in the x,y directions, will be significantly filtered out by averaging over the MRV.

The limited experimental and modeling evidence presented above shows that assumption (A) may hold after averaging the 3D fields of some SVA properties, while it remains to be demonstrated whether assumption (B) also holds, i.e. whether all relevant SVA properties may be measured directly over an entire MRV or multiples of it.

Discussion: Underlying assumptions and feasible observations

Soil

Eqs. 3 and 4 imply that soil structure and state may only change continuously with depth. For example, soil thermal and capillary conductivity depend on depth only. This implies several assumptions, e.g., that both K_D and K_w depend mainly on soil water content and that the latter changes only with depth. Soil type is taken into account by using different K_Ds and K_ws, and 3D conservation equations may be used to take into account macroscopic features of soil hydrology such as artificial drainage or the rooting pattern of row crops. Both the 1D or 3D equations, however, define variables and parameters by neglecting the inherent 3D structure of soils (see Figure 3) and assuming that measurements of W and K_w provide mutually consistent averages over an MRV.

At small length scales the soil structure is far from 1D (Figure 3), and size distribution of grains, aggregates and pores has an obvious impact on the actual spatial distribution of water, liquid and vapor, and of air. Water flow is a very complex process, which involves significant fluid – surface interactions in the pore space and phase transitions. Water flow as described by Eq. 4 is a totally different process. Water is assumed to move down the vertical gradient of ψ, assumed to account for all forces acting on water, with $K_w(z)$ accounting for all soil – water interactions determining the rate of displacement of both liquid and vapor in the soil space.

The state variables T_s and W are inherently 3D fields (see Figure 3), albeit at spatial scales comparable with the size of grains, pores and aggregates. As regards W, most experimental techniques, e.g. TDR or neutron scattering, directly provide averages over a rather large soil-sampling volume. The latter may or may not be significantly larger than the MRV depending on soil type and conditions (e.g. large aggregates, swelling/ shrinking cracks). Far less clear is whether the T_s measured by usual devices (thermoelectric transducers) is consistent with the T_s defined by Eq. 3. Moreover, heat flow is determined by the local temperature gradient and the extent of contact of soil particles, given the large difference between the thermal conductivity of air and of soil particles. The latter implies that streamlines for heat flow are closely correlated with soil structure. Methods (e.g. De Vries 1952) to estimate volume-averaged soil thermal conductivity from size and arrangements of soil particles have been developed, but it has been observed that differences between measurements and estimates of soil thermal conductivity increase with aggregates size (Hadas 1977). With increasing aggregates size, measurements relate to local heat-transfer processes in a complex 3D system and do not relate to soil thermal properties averaged over an MRV. For hydrological and climatic applications at a range of scales between fields to regions, the MRV properties are derived from soil maps in combination with pedo transfer functions, which introduces another source of error in the modeling concept.

Canopy space

Air temperature and vapor pressure as defined by Eqs. 5 and 6 can be easily measured by available devices. Spatial variability in the x,y directions of T_{ac} and e_{ac} may be neglected because of convective mixing over length scales comparable with the length scale of vegetation canopies. For moderately sparse canopies, the relaxation time of gradients in the canopy space is very short (Jia in press) and the measurements of Jacobs, Van Boxel and El-Kilani (1995) confirm that this assumption is acceptable. A second major assumption is that heat and vapor flow are proportional to the gradient through $K_h(z)$. This is unlikely to hold in the canopy space where multiple sources and sinks are present over the length scale of transport (size of eddies), as shown by, e.g., Legg and Monteith (1975), Denmead and Bradley (1985) and Finnigan and Raupach (1987). As noted by Dolman and Wallace (1991) and Raupach (1989) the concentration of scalars (temperature, water vapor, CO_2) in the canopy space is the sum of a flux–gradient term (far field) and a source–sink term (near field). Van den Hurk and McNaughton (1995) concluded that the difference in estimated bulk evaporation of a soil–vegetation system between the classical K-theory (flux–gradient) and the Lagrangian (source–sink) formulation was not large In hydrological and climate studies . within canopy heat transport is represented by a simple resistance type of approach, i.e. by assuming that the effect of distributed heat sources and sinks can be separated from temperature gradients.

Interestingly, Wilson et al. (2003) confirmed this finding, but emphasized that the two heat-transfer formulations led to large differences in simulated radiometric temperature of foliage. In other words, the processes determining the significant variability of observed foliage temperature (see Figure 4 on Color pages elsewhere in this book) are likely to be significantly different from processes described and variables defined by models such Eqs. 5 and 6. The latter implies that our observations may not relate to model variables in any simple way.

Atmospheric surface layer

The image data in Figure 5 point to a first striking difference between local measurements of air temperature and vapor pressure, i.e. at a specific location in the 10 km x 10 km domain, and T_a and e_a as defined by Eqs. 7 and 8. The MRV appears to be rather large and, therefore, local and instantaneous observations of T_a and e_a would be random samples of highly variable spatial fields, rather than constant at a given height as defined by Eqs 7 and 8. Moreover, the patterns shown in Figure 5 are also quite variable in time. The highly variable fields of T_a and e_a imply that free convection and (near field) radiative interactions are significant. The implication is that calculation of fluxes from measurements of scalar concentrations (T_a, e_a and CO_2) may not verify conservation of energy and mass. On the other hand, is not clear whether the average concentrations that verify the conservation principle in the simple form of Eqs. 7 and 8 can actually be measured.

The argument developed above would have required a broad and accurate review of literature. On the other hand it suggests that a solution to the well-known difficulties in modeling land – atmosphere exchanges must be sought through a significantly more precise understanding of the observations we use, rather than in the technical details of parameterizations. We are aware of the successful validation of many LSMs, but we contend that there is much to learn from a critical and comparative analysis of the nature of observed vs. model variables and parameters.

Approach to solution: modeling observations

To illustrate how a detailed characterization of (elements of) the SVA system can be used to understand better the nature of observations, we will analyse in detail the case of the surface temperature of a vegetation canopy (see Figure 4 on Color pages elsewhere in this book) in relation to modeling heat flow in the canopy space. A more detailed description of approach and results has been given by Jia (in press).

Two canopy constructions, 1D and 3D, will be used. The 3D construction will be applied to radiation-penetration calculation at any point in a canopy. The direct radiation flux at each point (considering a small surface at that point) is simply a fraction of solar radiation at the top of canopy with a proportion of penetration probability, while all the possible sources for diffuse radiation to a point must be taken into account. A one-dimensional scheme will be utilized for heat and water-vapor transfer in the canopy space. Though grid points (either leaves or soil surface) absorb different amount of net radiation depending on their locations in the canopy, heat and water-vapor transfer inside the canopy are controlled solely by the vertical variability of wind speed, air temperature and vapor pressure. This model structure preserves one essential feature of actual canopies: foliage temperature and the term S_H in Eq. 5 changes with the location (x,y,z) in the canopy space.

Jia (in press), following Welles and Norman (1991), defined a 3D vegetation canopy by means of sub-canopies whose outer envelope has an ellipsoidal shape. Sub-canopies can be individual trees or crop plants, even entire rows. Foliage within the ellipsoidal envelops of sub-canopies are assumed to be randomly distributed with any arbitrary function of zenith angle and random azimuthal distribution. This canopy construction can be used to compute the budget of direct, diffuse and emitted radiative fluxes at any point in the canopy and finally the leaf temperature, taking into account leaf orientation besides position in the canopy space. Neglecting the effect of photosynthesis and heat storage in the leaf, the energy balance for leaves in grid point $\vec{\mathbf{r}}$ and leaf angle class $\mathbf{\Omega_L}$ can be written as:

$$R_n(\vec{\mathbf{r}},\mathbf{\Omega_L}) = H_L(\vec{\mathbf{r}},\mathbf{\Omega_L}) + \lambda E_L(\vec{\mathbf{r}},\mathbf{\Omega_L}) \tag{13}$$

where $R_n(\vec{\mathbf{r}},\mathbf{\Omega_L})$ is net radiation flux density, $H_L(\vec{\mathbf{r}},\mathbf{\Omega_L})$ is the leaf sensible heat flux density, $\lambda E_L(\vec{\mathbf{r}},\mathbf{\Omega_L})$ is the leaf latent heat flux density, all in units of Watts per m^2 of leaf area in the leaf angle class Ω_L.

Radiometric observations of the foliage temperature can be used to estimate and model vegetation – atmosphere heat exchange (see e.g. Menenti 2000). It is often assumed that the observed radiometric temperature T_{rad} is equivalent to the radiometric temperature of a flat, uniform, infinite (leaf) surface. Experimental studies (e.g. Kustas et al. 1989; Beljaars and Holtslag 1991; Stewart et al. 1994; Troufleau et al. 1995) have shown that an empirical matching parameter, the so-called kB^{-1} (after Owen and Thomson 1963; Chamberlain 1968), can be estimated, although it depends on both vegetation type and environmental conditions. The model just outlined helps to understand the underlying cause.

The thermal heterogeneity, due to the location-dependent energy balance, within a vegetation canopy implies that the observed radiometric temperature of an MRV sample of the canopy changes with solar elevation, density of leaves and the angle distribution of leaves. The magnitude observed by a radiometer placed above a vegetation canopy is the top-of-canopy (TOC) brightness temperature (François, Ottlé and Prévot 1997) and is simply derived from the measured radiance R_λ using the inverse of Planck function. Therefore, the observation of surface brightness temperature and the observation of TIR radiance are equivalent. TIR radiance from a surface is usually referred to as 'exitance', i.e., TIR radiance emitted or reflected by the surface concerned.

Usually, the TOC brightness temperature is measured by a radiometer at certain channels (centered at some wavelengths) and in a particular direction (θ_v,ϕ_v) where θ_v is the zenith view angle and ϕ_v the azimuth view angle. The radiometer measures radiance with an instantaneous field-of-view (IFOV) Ω_v. The portions of canopy components with different surface temperatures in the IFOV of the radiometer will change with the view angles of the observation (Figure 6). As a consequence, strong anisotropy in exitance, i.e. a significant variation in surface brightness temperature with (θ_v,ϕ_v), can be observed over thermally heterogeneous systems, particularly for most sparse canopies. For instance, Kimes and Kirchner (1983) observed in a cotton field that the difference in radiative temperature between the 0° (mixture of vegetation and soil) and the 80° (vegetation only) zenith view angles was 16.2°C around noon, while the difference was only 0.9°C in the early morning. Lagouarde, Kerr and Brunet (1995) observed a difference of up to 3.5 K for a corn canopy and 1.5 K for grass (20 cm high) with a view zenith angle between 0° and 60° and around solar noon.

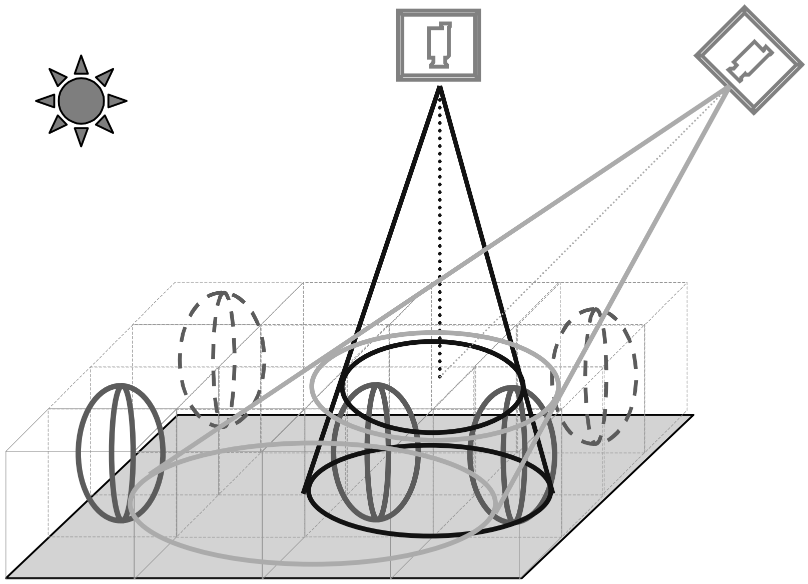

Figure 6. Observation of TOC brightness temperature at different view directions. The circles represent the footprints of IFOV, respectively at TOC and the bottom of the canopy at the respective view directions. The components in the volume between TOC and the bottom circles are observed by the radiometer above the canopy (adapted from Jia in press)

The radiance measured by a radiometer can be computed for any given illumination and view geometry and using a realistic 3D representation of the canopy. By taking into account the position of the radiometer relative to the canopy, we can also reproduce how the radiometer samples a 3D canopy. We will call this a complete model of observations of TOC brightness temperature and it leads to:

$$B[\lambda, T_{\mathrm{b}}(\theta_{\mathrm{v}}, \phi_{\mathrm{v}})] = \sum_{k=1}^{N_c} [f_k(\theta_v, \varphi_v) R_k(\lambda, T_k)] + \sum_{k=1}^{N_c} [f_k R'_k(\lambda)] + R^{\downarrow}_{\mathrm{atm}} \uparrow + R^{\downarrow}_{atm} \uparrow' \quad (14)$$

where B is the Planck function which relates surface brightness temperature to the TIR radiance, N_{c} is the number of grid points k (k=1,2,3......N_{c}) that would be seen by the sensor, $f_k(\theta_v, \varphi_v)$ is the fraction of the linear dimension of the volume δV of a grid point seen by the radiometer within the IFOV, $R_k(\lambda, T_k)$ is the radiance emitted by the grid point k to the sensor; $R_k'(\lambda)$ is the reflected radiance by the components in the kth grid point of radiance emitted by the components in the surrounding grid points, $R^{\downarrow}_{\mathrm{atm}} \uparrow$ is the reflected down-welling atmospheric TIR radiation by the components at all the grid points in the IFOV, and $R^{\downarrow}_{atm,k} \uparrow'$ is a term related to multiple scattering by canopy components of the down-welling atmospheric long-wave radiation. To predict the TOC T_{b0} observed in the case presented in Figure 4, we would need to characterize canopy geometry in detail.

From Eq. 14, it appears that the directionality of exitance from a canopy is a complex function of the radiance from components (a function of component temperatures and emissivities), the thermal radiation exchange between the components inside the IFOV and between the components inside the IFOV and those in the surroundings of the IFOV, and the canopy structure represented by $f_k(\theta_v, \phi_v)$. These factors play their roles in the relationship between the heterogeneity in TIR radiance and the anisotropy of exitance, i.e. of *the magnitude we actually measure*.

From the general case Eq. 14, simpler equations may be derived analytically to predict observations of simpler systems, as done by Jia (in press), who considered the cases of a foliage – soil system having four, two and one component. The hypotheses required to proceed from the complete to the simplest model were highlighted. Particularly, it was emphasized that there is no unambiguous way to interpret observations of a 3D, thermally heterogeneous canopy as the temperature of a homogeneous mixture of foliage and soil. This is due to the fact that we can define independently either the radiometric temperature or the emissivity of a 3D, thermally heterogeneous canopy, but not both of them.

To derive a simpler four-component model of T_{b0} at TOC we need to assume the following:

1) The soil and leaf surfaces are Lambertian. The sunlit and shadowed leaves have identical emissivity ε_f, and the sunlit and shadowed soil have identical emissivity ε_s;

2) Sunlit foliage, shadowed foliage, sunlit soil and shadowed soil have the respective effective temperatures T_{f_s}, T_{f_sh}, T_{s_s} and T_{s_sh}. The effective temperature for each component is defined as the ensemble temperature of all the respective components in the canopy by Planck law. Such relation is written as

$$B(T) = \sum_i f_i B(T_i) \tag{15}$$

where $T = T_{f\_s}, T_{f\_sh}, T_{s\_s}$ or T_{s_sh}; foliage is assumed consisting of finite facets with surface temperature T_i;

3) The canopy geometry is characterized by the fraction of each component area occupied in the IFOV – the component fractional cover. The component fractional covers in the IFOV change only with zenith view angles.

Additional hypotheses lead to equations to predict the observed TOC for a two- and single-component temperature. It remains to be verified by experiments that the simpler models may provide accurate predictions of the TOC T_{b0} or, in other words, that we may correctly predict radiance observed over a 3D thermally heterogeneous canopy without taking into account the full variability of foliage and soil temperature within the canopy.

For each of the cases above, a corresponding heat-transfer model can be established, as documented in literature, e.g. Choudhury and Monteith (1988), Friedl (1995), Anderson et al. (1997) and Chehbouni et al. (1996). The results of experiments and model analyses presented by Jia (in press) proved that:

(a) the soil and foliage temperatures of a two-component soil – foliage temperature can be determined from bi-angular observations of TOC T_{b0} ;

(b) the soil- and foliage-component temperatures can be used to estimate vegetation – atmosphere heat exchange with a dual source (i.e. soil and foliage) heat-transfer model.

Conclusions

The case of the radiometric temperature of vegetation canopies illustrates some of issues mentioned in the Introduction and Background sections. First, the detailed model of observations shows that the observed radiometric magnitude is quite different from the radiometric temperature defined and applied in models of the SVA system. Second, assumptions needed to arrive at the surface temperature as defined in models were identified. Third, the stepwise derivation of simpler models of observed radiance provides guidance towards simpler models of land – atmosphere heat exchange. The two-component model of observed radiance, for example, defines effective soil- and foliage-component temperatures, which can be used in dual-source models of heat exchange.

The example presented documents a general issue. Given the 3D nature of soil, of vegetation canopies and of the convective boundary layer, energy and mass transfer is determined by the spatial organization of object components, e.g. pores and grains in soil, and by their properties. Heat released by a vegetation canopy is determined by the position of leaves, whether they are sunlit or not, the amount of direct and diffuse radiation absorbed and convection of air in their vicinity. Integral measurements of say heat flux in the soil or exitance provide a measure of total energy exchanged.

We have shown that such observations can be modeled as a complex, non-linear function of state variables and properties of individual elements (leaves, soil). The combination of non-linearity and heterogeneity, as in the relatively simple case of the brightness temperature, implies that models, variables and parameters cannot be assumed to be all scale-invariant. If we choose a state variable to be scale-invariant (e.g. the brightness temperature), we need to derive an appropriate definition of other variables and parameters, emissivity in the case presented, from a detailed model of our integral (macroscopic) observations.

The latter is a common issue when dealing with observations of all three elements of the SVA system: soil, vegetation and atmosphere are all inherently 3D and actual processes occur and need to be understood at the inherent scale of variability. On the other hand the need remains to arrive at robust models of the SVA system at the continental and global scale. Use of global coverage, low spatial resolution observations has become widespread. The detailed analyses of the spatial organization of SVA elements and of energy- and mass-transfer processes at high spatial resolution outlined in this paper should be pursued towards replacing intuition-driven upscaling with a precise evaluation of the steps and assumptions involved when using observations, particularly from space, for this purpose.

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