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The Atmosphere-Vegetation-Soil System

1.1 Introduction

Whereas roughly 70% of Earth's surface is covered by oceans, the remaining 30% of land has a profound influence on processes in the atmosphere (e.g., differential heating, drag, evaporation and resulting cloud formation, composition of the atmosphere). This impact is due to the large variability in the properties (e.g., albedo, roughness, soil type, land cover type, vegetation cover) and states (e.g., soil moisture availability, snow cover) of the land surface. The processes occurring at the land surface are often grouped under the terms biogeophysical and biogeochemical processes (Levis, 2010): they influence the state and composition of the atmosphere both through physical and chemical processes, and biological processes play an important role in both.

Although the interface between Earth and atmosphere is located at the surface, subsurface processes in the soil are of major importance because part of the energy and water exchanged at the surface is extracted from or stored in the soil. Plants play an important role in extracting water from deeper soil layers and providing it to the atmosphere. In return, processes in the soil and plants (e.g., transport of water, solutes, and energy) are strongly influenced by atmospheric processes (e.g., evaporation and precipitation).

The interface between Earth and atmosphere is part of the continuum that ranges from the substrate underlying soils to the top of the atmosphere. The overarching subject of this book is the transport of energy, matter (water, solutes, CO₂), and momentum in the atmosphere-vegetation-soil continuum. In some cases this transport occurs within one of the compartments (e.g., redistribution of solutes in the soil); in other cases exchange over the interface between different compartments takes place (e.g., transpiration of water by plants).

In the context of this book we limit the extent of our subject both in the vertical direction (see Figure 1.1) and in the time scales considered. The lower boundary of the domain is located at that level in the soil where the *yearly* variation in temperature

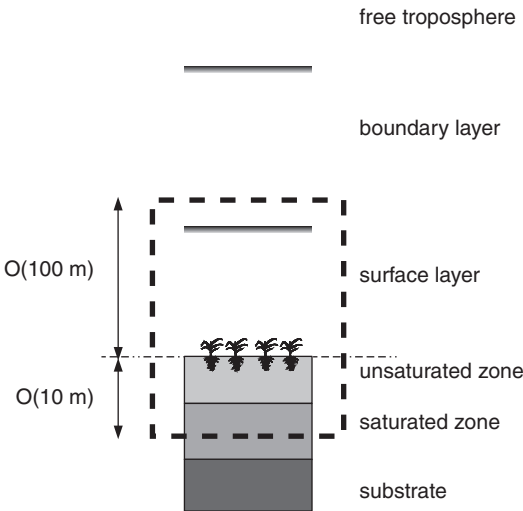


Figure 1.1 Various layers in the soil-vegetation-atmosphere continuum. The rectangular box indicates the vertical extent of the domain covered in this book. The vertical coordinate is roughly logarithmic.

and soil moisture has disappeared, whereas the top is located at the top of the surface layer, which occupies roughly the lower 10% of the atmospheric boundary layer (ABL, the layer where the *diurnal* cycle of surface heating affects the flow). The dynamics of the ABL itself are not part of this book, but are dealt with elsewhere (e.g., Stull, 1988; Vilà-Guerau de Arellano and van Heerwaarden, forthcoming).

The time scales considered range roughly from the diurnal cycle to the yearly cycle, although for the atmosphere turbulent fluctuations on time scales of less than a second are dealt with as well. As an example of variations on time scales from days to a year, Figure 1.2 shows daily rainfall fluxes and simulated daily evapotranspiration and drainage fluxes of a grass on a sandy soil in a Dutch climate and with a deep groundwater table. Although the rainfall fluxes are well distributed over the year, the evapotranspiration and drainage fluxes have a clear yearly pattern. Potential evapotranspiration follows the pattern of solar radiation. In the summer season with a high atmospheric demand, dry periods may cause a large difference between potential and actual transpiration. Drainage occurs mainly in winter periods with a low atmospheric demand and soil at field capacity. Compared to rainfall fluxes, drainage fluxes are much smaller and show a more gradual pattern.

Although the limitation of discussed time scales to a year is arbitrary, it suits the discussion of most processes well. That is not to say that no interesting and relevant processes occur at longer time scales, such as the interannual memory in vegetation cover (e.g., Philippon et al., 2007).

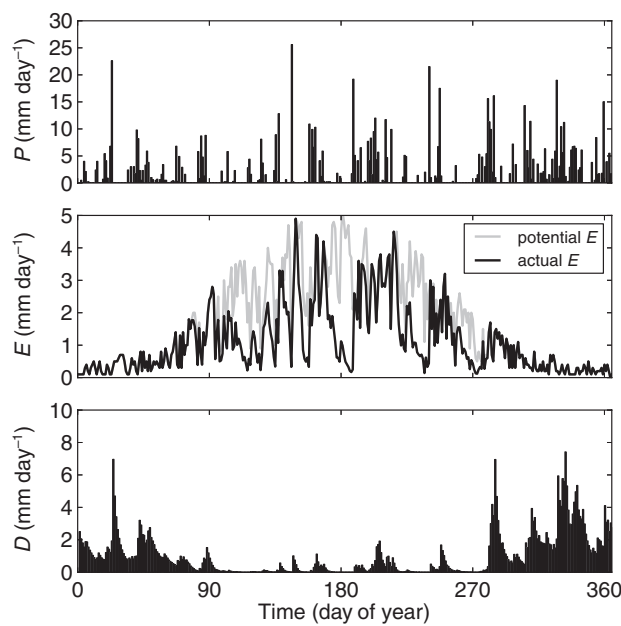


Figure 1.2 Daily rainfall fluxes (top) and simulated daily evapotranspiration (middle) and drainage fluxes (bottom) of grass on a sandy soil in a Dutch climate and with a deep groundwater table. Potential evapotranspiration in the middle panel is the evapotranspiration that would occur if sufficient water would have been available to the plants.

1.2 Conservation of Energy and Mass

To study the processes at Earth’s surface in a quantitative way we first need to define that interface clearly. This is done with the help of a control volume that contains the interface. Any difference between the inflow and outflow of a quantity will result in a change in the storage in the control volume. Formally, the conservation equation for an arbitrary quantity can be stated as:

$$\sum F_{\text{in}} - \sum F_{\text{out}} = \Delta S \tag{1.1}$$

where $\sum F_{\text{in}}$ and $\sum F_{\text{out}}$ are the summation of all fluxes into and out of the control volume, respectively, and ΔS is the change in storage (see Figure 1.3). To capture all important processes we start with a control volume that contains (part of) the soil column, as well as the part of the atmosphere into which the vegetation protrudes. In the horizontal direction the control volume has an arbitrary size, but horizontal homogeneity is assumed.

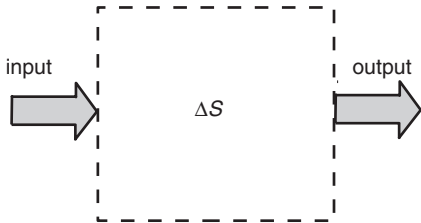


Figure 1.3 Concept of a control volume. Change in storage is due to an imbalance between input and output. Note that inputs and outputs can occur at any face of the control volume, not just the sides.

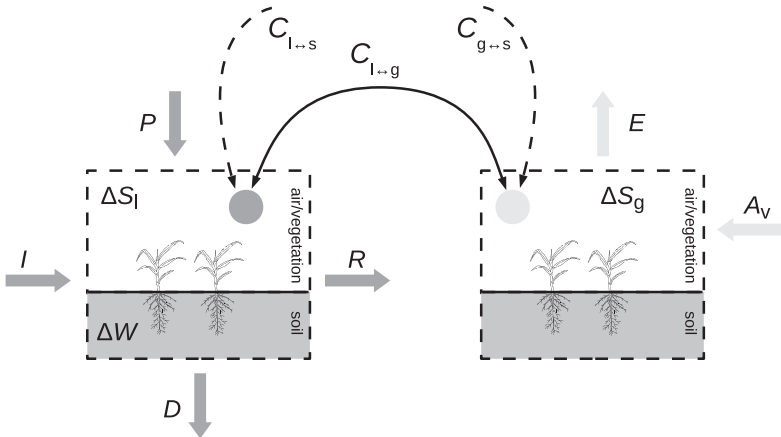


Figure 1.4 Control volume for water: liquid water (left) and water vapour (right). The direction of the arrows holds for typical daytime conditions. Dark grey arrows denote transport of liquid water, whereas light grey arrows are used for transport of water vapour. Arrows between the boxes signify phase changes (e.g., $C_{l \rightarrow g}$ is evaporation): molecules of water do not leave the control volume but only move from one phase to another. The dashed arrows with $C_{l \leftrightarrow s}$ and $C_{g \leftrightarrow s}$ have their other ends located in the control volume for solid water, not drawn here. Other symbols are explained in the text.

1.2.1 Water Balance

Figure 1.4 shows the water balance of the control volume.¹ Because under typical terrestrial conditions water occurs in all three phases (gas, liquid and solid), a distinction has been made between water in the liquid phase and water in the gas phase (ice and snow have been discarded for simplicity, but could be accounted for analogously;

¹ Note that it is implicitly assumed that the control volume has a horizontal extent of one square meter. As a result all fluxes and storage terms, as used in Eqs. (1.2) and (1.3) should be interpreted as fluxes per 1 m² of ground area. This assumption is not a physical necessity, but simplifies the transition to the other chapters where usually fluxes are interpreted as flux densities (i.e., per unit area).

see Question 1.1). As water can change phase within the volume, all phases have to be accounted for in the total water balance of the control volume. The water balance then reads:

$$P + I - R + A_v - E - D = \Delta W + \Delta S_l + \Delta S_g \tag{1.2}$$

where we distinguish the following transports across the boundaries of the control volume: P is precipitation, I is irrigation (artificial supply of water), R is runoff, D is the drainage rate towards deeper soil layers, A_v is advection of water vapour (which can be positive or negative) and E is the water that leaves the system in gas phase (water vapour; see Section 1.2.3). The different inputs and outputs do not necessarily balance, so that water may be stored in the soil (ΔW , change in soil moisture content), on the soil or on the vegetation in the liquid phase (ΔS_l , e.g., intercepted rain or dew) and in the air (ΔS_g) in the gas phase.

Water molecules can change phase. In Figure 1.4 this is indicated with the phase change terms $C_{l \leftrightarrow g}$ (between liquid and vapour), $C_{l \leftrightarrow s}$ (between liquid and solid) and $C_{g \leftrightarrow s}$ (between vapour and solid). These phase change terms, however, do not occur in Eq. (1.2) because they do not change the number of molecules of water within the control volume but only the phase in which they occur. On the other hand, as energy is released or used when water changes phase, the phase change terms do affect the energy balance, as we see later.

- Question 1.1:** Figure 1.4 shows the mass balances for water of a control volume extending from the soil into the atmosphere, above the vegetation. Only liquid and gaseous water is dealt with.
- a. Sketch the mass balance for solid water.
 - b. Enumerate all interactions between the mass balances for each of the phases (i.e., which phase changes can occur that exchange water in one phase for another?).

1.2.2 Energy Balance

Whereas in the water balance we have to distinguish between the three phases of water, in the energy balance we distinguish between two forms of energy: sensible heat and latent heat. Sensible heat is the energy contained in a substance that can be extracted by cooling it. On the other hand, latent heat can be extracted only by a phase change. It could be considered as similar to potential energy: a ball on a hill has potential energy that could be extracted when it rolls down the hill. In a similar way, water vapour contains latent heat that would be freed if it condenses to liquid water. In the current context it is customary (but not necessary) to use liquid water as the reference (i.e., ice contains negative latent heat because it would require *input* of heat to bring it to liquid water).

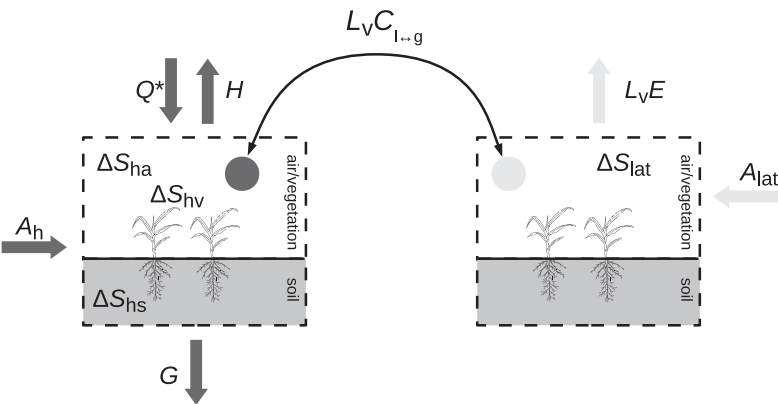


Figure 1.5 Control volume for energy: sensible heat (left, dark grey arrows) and latent heat (right, light grey arrows). Energy is exchanged between sensible and latent heat if water changes phase (here only phase changes between liquid and gas phase are considered: $L_v C_{l \leftrightarrow g}$). Storage of energy in the form of chemical energy in the plants (due to assimilation) as well as some other terms (see text) have been discarded. The direction of the arrows holds for typical daytime conditions.

The control volumes for sensible and latent heat are depicted in Figure 1.5. The complete energy balance equation corresponding to this figure is as follows:

$$Q^* - H - L_v E - G + A_h + A_{lat} = \Delta S_{ha} + \Delta S_{hv} + \Delta S_{hs} + \Delta S_{lat} \tag{1.3}$$

where Q^* is the net radiation (see Chapter 2), H is the sensible heat flux, G is the soil heat flux at the bottom of the control volume and $L_v E$ is the latent heat flux (where L_v is the latent heat of vaporization). If the inputs and outputs do not balance, heat can be stored in the air (ΔS_{ha}), in the vegetation (ΔS_{hv} , i.e., the vegetation becomes warmer) and in the soil (ΔS_{hs}). Finally, A_h and A_{lat} are the net advections of sensible and latent heat.

The latent heat flux plays a special role in the energy balance in the sense that it transports energy through the transport of water vapour. The actual energy consumption related to *evaporation* is contained in the term $L_v C_{l \leftrightarrow g}$, but because this term is internal to the control volume, it does not appear in the energy balance. The energy related to evaporation leaves the control volume as latent heat. In the case that there is a change in the water vapour content in the control volume (i.e., $\Delta S_{lat} \neq 0$) or non-zero advection of water vapour, the transport of latent heat *out of* the control volume ($L_v E$) may be unequal to the energy related to the phase change *within* the volume. Note that the opposite phase change can happen as well: when dew is formed water is transformed from the gas phase to the liquid phase and energy is released.

Some terms have not been accounted for in Eq. (1.3). Part of the solar radiation that hits the vegetation will be used for photosynthesis. This leads to a conversion of radiative energy to the storage of chemical energy in the plant material. This storage term

has been discarded (it is of the order of $10\text{--}25\text{ W m}^{-2}$ at midday, depending on insolation and the type of vegetation; see Meyers and Hollinger, 2004). Another term in the energy balance, seldom taken into account, is the sensible heat related to the input of precipitation into the control volume. If, for instance, during daytime conditions the temperature of the rain is lower than that of the air that it replaces in the control volume, a net exchange of energy out of the volume occurs. Further, if the cold rain-water percolates into the soil, a significant redistribution of energy within the control volume can occur (see Kollet et al., 2009).

1.2.3 The Link: Evapotranspiration

From Eqs. (1.2) and (1.3) the link between the water balance and energy balance becomes clear: the evaporation appears as a transport of mass in the water balance and as a transport of energy in the energy balance. The total water vapour flux that leaves the system is made up of a number of fluxes *within* the system: soil *evaporation* (E_{soil}), *transpiration* by the plants (T), and *evaporation* of intercepted water (E_{int}). Both transpiration and soil evaporation extract water from the soil subsystem and release it in the air subsystem, whereas in the case of interception the soil is bypassed. The sum of these three terms is called *evapotranspiration* and denoted by E (in the hydrological literature a commonly used symbol is ET).

Evaporation and transpiration occur simultaneously and there is no easy way of distinguishing between the two processes. Apart from water availability in the top soil, the evaporation from a cropped (or more general: vegetated) soil is determined mainly by the fraction of solar radiation reaching the soil surface. This fraction decreases over the growing period as the crop develops and the crop canopy shades more and more of the ground area. When the crop is small, water is lost predominantly by soil evaporation, but once the crop is well developed and completely covers the soil, transpiration becomes the main process. In Figure 1.6 the partitioning of evapotranspiration into soil evaporation and transpiration is plotted in correspondence to leaf area per unit soil surface below it (LAI). At sowing, nearly 100% of E comes from soil evaporation, whereas at full crop cover more than 90% of E comes from transpiration (Allen et al., 1998).

Apart from the direct link between water balance and energy balance through the occurrence of evapotranspiration in both balances, there is also a more indirect link. The two balance equations exactly stand for the two requirements needed for evapotranspiration: water should be available to be evaporated, and energy (through radiation) should be available to actually let the evaporation happen (see also Chapter 7). The availability of water in the soil in turn will largely be determined by the amount of rainfall. These two limiting factors – radiation and rainfall – translate into different behaviour of evapotranspiration in different regions of the world: in regions of abundant rainfall (relative to the evaporative loss) evapotranspiration correlates highly

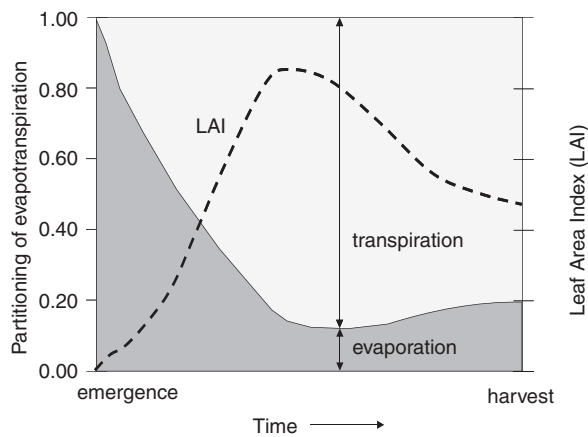


Figure 1.6 The partitioning of evapotranspiration into soil evaporation and transpiration over the growing period of an annual field crop.

with radiation, whereas in more arid regions radiation is not limiting, and evapotranspiration correlates with rainfall (Teuling et al., 2009). Although these correlations sketch the main picture on a seasonal or longer time scale, many regions may show different behaviour on shorter time scales. For example, mid-latitude regions that show a strong correlation between radiation and evapotranspiration may show a stronger dependence on soil moisture and rainfall after a prolonged drought (Teuling et al., 2010 and Chapter 8). On the other hand, semi-arid regions in which the seasonal pattern of evapotranspiration follows the seasonal pattern of rainfall may show a clear correlation of evapotranspiration with radiation *within* the rainy season (e.g., Schüttemeyer et al., 2007).

- Question 1.2:** Evapotranspiration is a combination of various fluxes: evaporation from intercepted water, soil evaporation and transpiration.
- a. What is roughly the impact of each of the terms on the soil moisture content at various depths?
 - b. At which location is the energy, needed for the phase change from liquid to water vapour, absorbed for each of the fluxes?

1.2.4 Simplified Balances

In many applications the control volumes presented earlier are vertically compressed to become a control surface (see Figure 1.7). Because a surface has no volume, storage terms will disappear. Besides, horizontal advection will vanish as well. Whereas in Eqs. (1.2) and (1.3) all fluxes occurred at the boundaries of the control volume (i.e., at a certain height above the ground or at a certain depth below the surface) now

Conservation of Energy and Mass

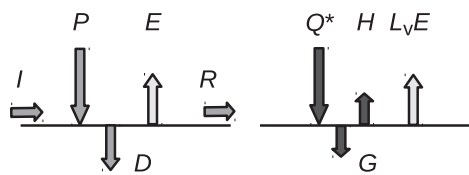


Figure 1.7 Simplified surface water balance (left) and energy balance (right). Very light grey arrows indicate transport of water vapour, light grey arrows show liquid water transport and dark grey arrows indicate energy transport.

all fluxes are supposed to occur *at* the surface: they are called *surface fluxes*. With this redefinition of the fluxes, the surface water balance and surface energy balance become:

$$P + I - R - E - D = 0 \tag{1.4}$$

$$Q^* - H - L_v E - G = 0 \tag{1.5}$$

Although these balance equations are appealing in their simplicity, the compression of the control volume to a surface may lead to problems in the interpretation of observed fluxes. In practice most fluxes will be observed at some height above a canopy, or at some depth below the surface (rather than at the hypothetical surface for which they are supposed to be representative). The omission of storage and advection terms in Eqs. (1.4) and (1.5) may cause a non-closure of the observed water balance or energy balance (i.e., the terms do not add up to zero). To solve this problem one should revert to the full Eqs. (1.2) and (1.3).

Note that the sign convention in Eqs. (1.4) and (1.5) is such that P , I and Q^* are considered as inputs, taken positive when directed towards the surface, whereas the other terms are considered as outputs, being positive when directed away from the surface. This sign convention is often used but arbitrary and other choices are used as well.

A final remark relates to the units of the transports in Eqs. (1.4) and (1.5). Rather than delineating a real surface with a given extent, those equations are usually applied to a unit surface area (e.g., one square meter). This implies that the units of the transport terms in the water balance are either volume per unit area per unit time (volume flux density) or mass per unit area per unit time (mass flux density). The terms in the energy balance have units of energy per unit area per unit time (energy flux density). Often, the terminology is used loosely: the term “density” is dropped and the word “flux” is used where a “flux density” is meant.

To summarize, the various terms in the surface water balance and energy balance are the subject of this book. But the processes are not studied only in isolation, but the interactions between them are at least as important.

1.3 Modes of Transport of Energy and Mass

In the presentation of the surface water and energy balances a number of transport processes have been introduced, without detailing by what means the transport takes place. For the transport of energy, three modes of transport are possible:

- Radiation (transport by propagation of electromagnetic radiation; no matter is needed)
- Conduction (transport of energy through matter, by molecular interactions; matter is needed, but the matter does not move [macroscopically])
- Advection² (transport of energy by the movement of energy-containing matter)

In the context of the surface energy balance, all three modes occur: net radiation is radiative transport, the soil heat flux is based on conduction, and the sensible flux is an example of advection (in the sense that turbulent transport involves the motion of energy-containing air).

For the transport of matter (e.g., water or solutes) the two modes of transport are

- Molecular diffusion
- Advection

In the present context, mainly advection is important, but it can have a number of different manifestations. In soil, water flows more or less smoothly, whereas in the atmospheric surface layer transport of matter (e.g., water vapour or CO₂) takes place by turbulence, where air containing the given constituent is moved from one place to another (see Chapter 3). Molecular diffusion plays a role in solute transport in the soil, and in the atmosphere in thin layers adjacent to surfaces (e.g., leaves).

One of the main objectives of this book is to quantify the various fluxes in water balance and energy balance. A generally used method to describe the flux of a quantity is based on an analogy with transport by diffusion on the molecular scale, that is, Fick’s law for diffusion of matter and Fourier’s law for heat diffusion:

$$F_a = -k_a \frac{\partial C_a}{\partial x} \tag{1.6}$$

where F_a is the flux density of quantity a in the x -direction, C_a is the concentration of a and k_a is the molecular diffusion coefficient for quantity a (k_a has units m² s⁻¹). Essential in the case of molecular transport is that k_a is known and depends only on the fluid or solid under consideration and on the state of that fluid or solid (temperature, pressure). For many combinations of transported quantities (e.g., momentum, heat, water vapour) and fluids (e.g., air or water), k_a is known and tabulated.

² Here the word ‘advection’ is used in the sense of large-scale (relative to the molecular scale) motion of matter. Sometimes the term convection is used for this as well, but this may cause confusion with thermal convection, and in some applications convection is the sum of advection and diffusion.