6. Land Surface Schemes and Climate Models

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> *Sometimes model equations are presented that make one wonder whether Nature knows them too.*

6.1 INTRODUCTION: CLIMATE MODELS

Computer models used for weather and climate prediction describe multiple processes simultaneously at a wide range of spatial and temporal scales. Individual synoptic pressure systems have a life time of ~10 days and a spatial scale of ~1000 km, whereas individual rain showers with a dimension of ~1 km exist for only ~1 hour. The spatial grids of typical climate models are able to resolve the synoptic systems, but necessarily treat processes at a smaller scale using empirical parameterizations.

The constitution of the atmosphere can essentially be described using five basic equations:

(a) *the equation of state*, also denoted as the gas law, that relates pressure, temperature and air density;
(b) *the equation of motion*, that describes flow in the presence of pressure gradients and buoyancy;
(c) *the mass conservation equation*, also known as *the continuity equation*, that will change the pressure of a volume when the amount of mass entering the volume is not equal to the mass leaving the volume;
(d) *the moisture conservation equation*, that keeps track of the various phases of water; and
(e) *the energy conservation equation* that describes the temperature change as a result of latent heat release, radiation absorption or mixing.

In climate models, these equations are solved to calculate the evolution of so-called *state variables* (temperature, water vapour, wind speed in two directions, and sometimes also liquid and/or ice-phase water, or trace gas concentrations) on the model grid. General Circulation Models (also known as Global Climate Models, GCMs) have a typical grid spacing of 2° in latitude/longitude direction, and about 30 vertical levels. In a GCM the globe thus consists of ~500 000 grid points, where calculations are carried out at ~30 minute intervals.

Processes not resolved at this spatial resolution are addressed using a set of *parameterizations*, often denoted by *the physical package*. A parameterization describes a process (e.g. precipitation) as a function of the state variables (e.g. temperature and (relative) humidity) using a (semi-)empirical equation. These parameterization equations usually contain some first principle physics (e.g. the thermodynamic description of air parcels), some empirical relationships (rain droplets form whenever the relative humidity exceeds a certain threshold), and some tunable coefficients (the saturation threshold that relates cloud fraction to relative humidity, or an assumed number of condensation nuclei).

Generally, the following groups of parameterized processes may be discerned, all acting on temporal and spatial scales smaller than resolved by the model grid:

(a) *Radiation*, where transmission through or absorption by the atmosphere, reflection or scattering of shortwave (solar) radiation by clouds, and longwave (infrared) emission by the air and surface are the key phenomena.
(b) *Condensation/evaporation*, where heat exchange by water phase changes gives rise to temperature changes of air parcels.
(c) *Convection*, where organized vertical motion of air is induced by heating or cooling layers in the surface–atmosphere column.
(d) *Vertical mixing by turbulence*, where wind shear and heating create small-scale eddies that mix air parcels efficiently.
(e) *Land surface processes*, where the presence of vegetation or ice controls the hydrological and radiative fluxes between the land surface and the atmosphere.
(f) *Gravity wave drag*, where breaking waves in the flow cause a net transfer of kinetic or potential energy.
(g) *Atmospheric chemistry*, where aerosols may develop or wash out and where chemical reactions change the composition of the atmosphere, with sometimes a large impact on the radiative properties of the atmosphere.

Climate model simulations are fairly sensitive to the formulation of these parameterized processes. A first-order property of climate models applied in studies addressing the effects of increased greenhouse gas emissions is the so-called *climate sensitivity*, i.e. the change of the global mean surface temperature in response to a doubling of the CO_2 concentration. This temperature response is a result of a chain of parameterized processes (radiative absorption by greenhouse gases, formation of clouds, response of vegetation transpiration under elevated CO_2 levels, etc.). Therefore, the development, calibration and validation of these parameterizations has deserved a lot of attention in the scientific literature.

A good introduction to the design, components and application of GCMs can be found in Trenberth (1992) and Holton (1972). In the remaining part of this chapter we will zoom in on the parameterization of land surface processes in climate models. First the processes that must be covered by these schemes are described, and later an overview is given of the general structure of land schemes, their calibration and validation. The final section addresses some topics that deserve increasing attention in present-day research.

6.2 THE FUNCTIONS OF LAND SURFACE SCHEMES

Land surface models (or LSMs) represent the exchange processes at the land surface–atmosphere interface, and the change of the state of the soil, snow pack and land ice. In essence, an LSM has the following tasks:

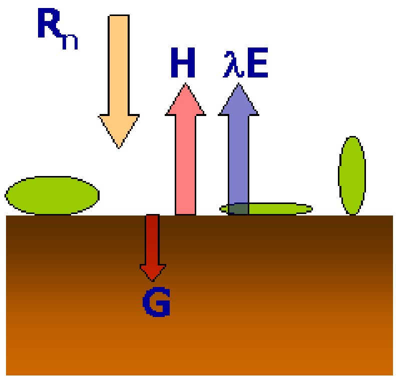

Fig. 6.1 *Schematic layout of the surface energy balance equation.*

(a) *Radiative reflection and emission*: The surface albedo varies with the type of surface. Dark forests reflect only ~10% of the incoming shortwave radiation, whereas fresh snow has an albedo of ~85%. Dry sand reflects more (~30%) than wet soil (~15%). Thus for a downward shortwave irradiance of 800 W m^{-2}, the surface reflects 80 to 720 W m^{-2}, depending on the type of surface. Longwave radiation is emitted as a function of the surface temperature and the longwave emissivity, ε. The latter quantity varies between 0.90 for dry sand to 1.00 for wet vegetation. The amount of longwave radiation emitted varies by ~4.5 W m^{-2} per degree K.

(b) *Partitioning of energy*: The net radiation at the surface R_n consists of incoming and reflected shortwave radiation $K^{\downarrow}$ and $K^{\uparrow}$, and incoming and emitted longwave radiation $L^{\downarrow}$ and $L^{\uparrow}$. R_n provides the main energy source in the surface energy balance equation, which reads:

$$R_n = H + \lambda E + G + F \tag{6.1}$$

and is shown schematically in Fig. 6.1

(c) R_n is thus partitioned over the sensible heat flux H (warming the air close to the surface during daytime), latent heat flux via evaporation λE, heat storage in the soil, snow pack and canopy G, and a small amount used for photosynthesis in plants F. The latter term is often ignored in LSMs. The partitioning over H, λE and G depends highly on the surface properties. In the absence of (soil) water λE is zero, whereas well-watered crops may use nearly all the net radiation for latent heat exchange. Insulating moss layers may reduce G to nearly zero, whereas in the absence of turbulence both H and λE approach zero and all net radiation may be compensated by G.

(d) *Partitioning of water*: The water balance equation of the land surface (shown schematically in Fig. 6.2) reads:

$$\frac{\partial (W + S + I)}{\partial t} = P - E - Q_s - Q_d \tag{6.2}$$

Here W is the total water content of the soil, S the snow pack and I the interception

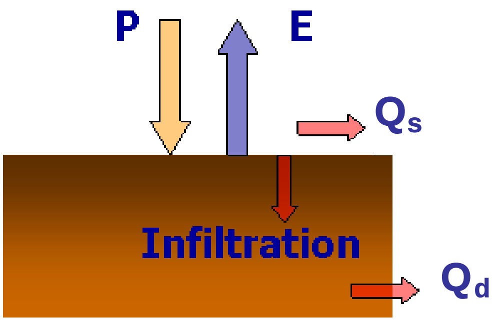

Fig. 6.2 *Schematic layout of the water partitioning at the land surface.*

layer on leaves and soil. These three terms together define the total amount of water stored in or on the soil, also known as the *terrestrial water reservoir* (expressed in m). This state is affected by the fluxes (m s^{-1}) on the right-hand side of equation (6.2). The fate of precipitation P depends highly on the land surface characteristics. Solid rock will be wetted by rain, and this wetness will eventually be evaporated (E), but the dominant part of the rain will be removed almost instantaneously as surface runoff Q_s. In vegetation covered areas the portion of rain that is intercepted by the leaves varies with the leaf area and intensity and the duration of the precipitation event. As much as 25% of all precipitation is intercepted in some tropical forest areas, whereas interception in sparsely vegetated vineyards is low. The remaining liquid precipitation infiltrates into the soil, where it is either stored in the soil water reservoir W, or taken up by vegetation roots, or it percolates to deeper groundwater tables (Q_d). In the absence of vegetation, evaporation is restricted to the bare soil surface, and depends on a water flux to the surface from below, and hence the water transport capacity of the soil. Vegetation with roots has a greater capacity to extract water from the soil for evaporation, but rooting profiles are dependent on the kind of vegetation. Snow precipitation does not percolate, but is stored on top of the surface where it either evaporates, or melts (M) in which case it is infiltrated into the soil or removed as surface runoff.

(e) *State evolution*: The state of the land surface is diagnosed from a series of evolving variables, known as the *soil state variables*: soil temperature, soil moisture content, and in some LSMs also the temperature of the ice pack, a separation between frozen and liquid soil water, or various stores of carbon. Like the state variables in the atmospheric model, these variables are *prognostic variables*, whose values change as a result of processes acting on them: e.g. transport of heat or water, thermodynamic phase change, extraction of water by vegetation, etc. (see equation (6.2) for an example of a state equation). The land surface model keeps track of these changes, and at any moment provides an estimate of the condition of the soil and surface. Model variables that are parameterized (e.g. fluxes) or derived from other variables (such as the value of the temperature at 2 m height obtained via interpolation between the surface and the lowest atmospheric level) are called *diagnostic variables*.

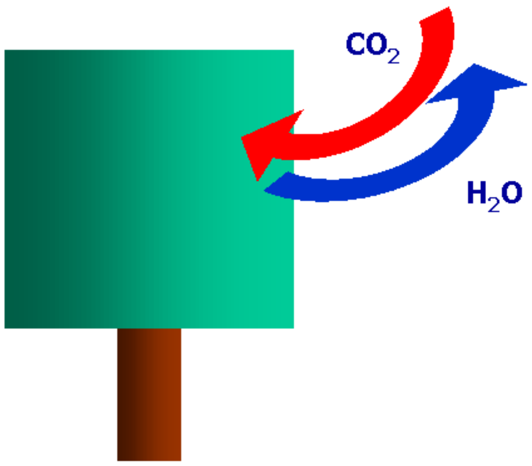

Fig. 6.3 *Schematic layout of the coupling between CO_2 exchange and canopy transpiration.*

(f) *Net carbon exchange*: A final function of land surface models to mention here is that they should represent the exchange of CO_2 between the land surface and the atmosphere. Not all land surface models take this process in to account, and not all land surface models treat the fate of the exchanged CO_2 similarly. In the simplest form, CO_2 exchange is treated in conjunction with water vapour exchange via plant leaves (Fig. 6.3), coupling the processes of photosynthesis and canopy transpiration through stomata. In more advanced models, CO_2 exchange is complemented by equations expressing the accumulation or decay of living biomass, and complex schemes exist that partition this biomass over roots, stems, leaves and storage organs, depending on plant phenology, nutrient and water availability, etc.

6.3 GENERAL STRUCTURE OF LAND SURFACE SCHEMES

In general, land surface schemes in climate and weather prediction models are designed around the basic conservation equations of energy (equation (6.1)) and water (equation (6.2)). During each time step the following series of procedures is followed. They are listed here in an order starting with the state variables we are finally interested in, and gradually zooming in on the components that have to be calculated first. Computer code normally follows the route in reverse order, i.e. starting with Section 6.3.6 and gradually progressing to finally, the update of the state variables, Section 6.3.1 .

6.3.1 Update of the state variables

The final step in a land surface scheme is an update of the state variables. For *soil temperature* T_s, the prognostic equation to be solved is given by:

$$\rho C_s \frac{\partial T_s}{\partial t} = -\frac{\partial G}{\partial z} \qquad (6.3a)$$

where G is the vertical flux of heat, and ρC_s is the volumetric heat capacity of the soil. At the surface, G is defined as a residual from the energy balance equation (see below). In deeper layers, G is defined as:

$$G = -D_T \frac{\partial T_s}{\partial z} \tag{6.3b}$$

For soils with a uniform vertical distribution of ρC_s and thermal diffusivity D_T, an analytical solution can be given for the evolution of T_s when a sinusoidal forcing of a surface temperature is provided. This analytical solution is, in some land surface schemes, implemented as a so-called *force-restore* equation, where heat exchange between two or more layers is treated as a combination of a forcing from the surface and a damping (restoring) term provided from the slowly responding deeper layer(s). However, most LSMs use a diffusion equation as formulated above, instead, which gives better results when the surface forcing has rapid changes and/or the vertical distribution of thermal properties of the soil varies.

Concerning the *soil water equation*, modellers with a meteorological background tend to follow the treatment of soil temperature, and an equivalent diffusion equation is defined, reading (with $W = \int_{d_s} \theta \mathrm{d}z$, d_s being the total soil depth and θ the volumetric soil moisture content):

$$\rho_w \frac{\partial \theta}{\partial t} = -\frac{\partial F_w}{\partial z} - \rho_w S_R \tag{6.4a}$$

with F_w the water flux, which within the soil is given by:

$$F_w = -\rho_w \left[D_w \frac{\partial \theta}{\partial z} - K \right] \tag{6.4b}$$

Hydraulic conductivity K and diffusivity D_w are discussed in Section 6.3.3. At the surface the forcing flux F_w consists of the sum of infiltrating precipitation and melting snow minus bare ground evaporation, and the bottom boundary condition is a parameterized value of the deep drainage flux (see below). Water exchange takes place between the various discrete soil layers due to a diffusion-like process over a hydraulic pressure gradient and gravity drainage. Root extraction (S_R) takes place using a specified distribution over the soil, and is equal to the total canopy transpiration rate. Using equation (6.4) climate models naively apply the Richards equation derived for small-scale exchange processes at the scale of a model grid box, ignoring the many non-linear effects of the hydraulic properties, or incorporating these nonlinearities in so-called *effective* hydraulic parameters. Similar to the temperature solution, some models carry a force restore equation for this variable.

However, alternative formulations exist that are designed from a more hydrological point of view. Soil water content is not necessarily considered as a vertical stack of layers, but is defined using a water holding capacity that varies across the grid box. As such, soil water content is treated in relation to the orographic organization of water in subcatchments (see Fig. 6.4).

Simple equations can be derived for the evolution of the *snow volume* (usually expressed in liquid water equivalent units: accumulation via frozen precipitation, decay

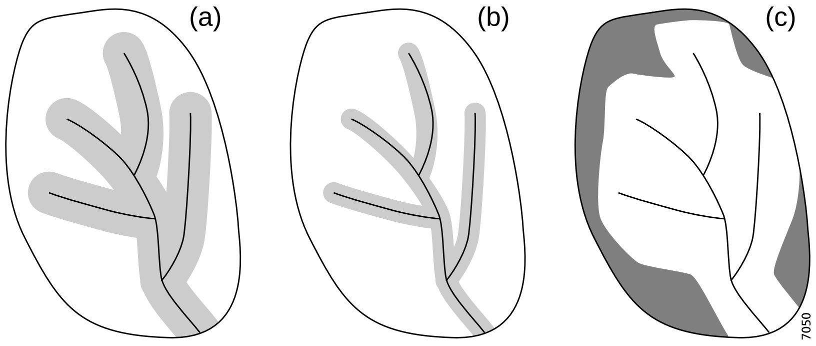

Fig. 6.4 *Representation of saturated fraction of grid box varying over time; black line: river network; grey area in (a) and (b) indicates saturated fraction; grey area in (c) represents area below wilting point (after Koster et al., 2000)*

due to melting or sublimation), but the complexity of the *snow temperature* evolution varies widely across models. Simple schemes treat the snow pack as a single layer with a dynamic heat capacity (which is a function of the total snow mass), but complicated multi-layer models also exist that elegantly treat melting due to absorption of radiation in the top of the snow pack, vertical percolation of liquid water through the snow volume, and refreezing depending on ambient conditions.

Two other states of the snow volume are often treated using a prognostic equation: the *snow albedo* and the *snow density*. Snow albedo gradually reduces in time due to dust accumulation on the surface and compaction of the snow particles, and fresh snow fall resets the snow albedo to higher values. Snow density increases in time owing to compaction, surface melt and deeper refreezing, and this variable is necessary to translate the snow depth measured in water equivalent units into physical snow depth which can be compared with field observations. In addition, the snow density affects the heat capacity of snow (ρC_{sn}), which has a large influence on the time and depth evolution of the snow temperature. Vertical temperature transport in multi-layer snow models is dependent on the snow density profile, and in these models the timing of the snow melt may thus be dependent on the snow density evolution.

Similar to the snow volume, a simple budget equation is used to express the evolution of the amount of *intercepted water*. This (small) reservoir is filled with precipitation, and vanishes again as a result of evaporation. The maximum amount of water that can be intercepted greatly depends on the leaf area index: multiple leaf layers per unit surface area can hold more water in the interception reservoir than a plain bare soil.

6.3.2 Calculation of the hydrological and thermal fluxes in the soil

The water fluxes acting on the soil water content consist of the following terms (see also equation (6.2)):

(a) *Precipitation* can be considered as a forcing to the land surface scheme. Some schemes use a temperature threshold to discriminate between rainfall and snowfall.

(b) *Interception* is the part of the precipitation flux that is collected on leaves and the ground surface. It is equal to the precipitation rate until the interception reservoir is full, and zero afterwards.
(c) The portion of rainfall not intercepted or stored in the snowpack is called the *throughfall rate*. It is part of the flux of water entering the soil, called the *infiltration rate*.
(d) Another component of the infiltration rate is the *snowmelt,* governed by the snow pack temperature.
(e) Within the soil, water normally only moves vertically. This vertical transport consists of a *diffusion-like flux* over a gradient of the hydraulic head (see Section 6.3.3 below), and a *drainage rate*, which is governed by the equilibrium between the gravity force and the friction force of the soil matrix acting on water. (The soil matrix consists of the soil particles and pores partially filled with water).
(f) Also within the soil, water is removed by *root extraction,* which may vary with depth and equals the total canopy transpiration.
(g) The *evaporation flux* consists of the canopy transpiration, evaporation from the interception reservoir, evaporation from the upper bare ground layer, and evaporation from snow (sublimation). These terms are detailed in the energy budget, Section 6.3.4 below.
(h) In layered land surface schemes the *runoff* consists of three terms:
    i. The *drainage* through the bottom of the lowest soil layer (equal to the lowest layer hydraulic conductivity – see Section 6.3.3 below); this normally is a slowly responding term in the soil hydrological budget, owing to the time needed to propagate the infiltration through the soil column.
    ii. The *surface runoff*, which results from an infiltration rate exceeding the maximum uptake capacity of the soil; it responds immediately to the infiltration signal, and thus to the precipitation variability
    iii. The *interflow* is runoff generated in intermediate layers, where via diffusion and drainage the soil water content exceeds the saturation value.

Land surface schemes with a more hydrological design instead of a layered structure, have similar notions of various contributions to the runoff, albeit that the explicit formulation of runoff generating sub-reservoirs allows more flexibility (and adjustment) of various time scales in the runoff generation process.

The soil thermal fluxes are simply calculated by solving a standard diffusion equation (equation (6.3)) with a thermal diffusion coefficient D_T that in some LSMs depends on the soil water content. The occurrence of *soil freezing and melting* is a special case. A considerable amount of latent heat is released when moist soils freeze, and this latent heat is used again during the melting phase. This causes a so-called *zero-degree barrier*: the seasonal evolution of soil temperature holds a while around 0°C, both during the freezing and during the melting stages. In many LSMs this effect is parameterized by accounting for the latent heat release associated with the water phase change added as an extra term to the specific heat content of soil. This term is zero when the soil temperature is strongly different from 0°C and has higher values around the freezing point.

6.3.3 Calculation of the soil exchange coefficients

Prior to calculation of the hydraulic and thermal fluxes in the soil, the exchange coefficients in equations (6.3) and (6.4) must be specified. This is not a trivial task, since – as indicated above – the value of these coefficients must represent the physical processes at the scale of operation, which is usually much larger than the values tabulated in many studies based on field investigations. For the soil thermal diffusivity D_T, various parameterizations exist, and they roughly fall into categories that assume fixed values depending on soil texture only, values that vary with the soil moisture content, or a blend of the two.

For the soil hydraulic coefficients K (conductivity) and D_w (diffusivity) various "schools" of approach are used in the meteorological land surface community. Many models express K as function of soil water content θ, a conductivity for saturated soils $K_{,s}$ that depends on the soil texture, and a number of empirical shape parameters. The formulation of Clapp & Hornberger (1979) is widely used:

$$K_w = K_s \left(\frac{\theta}{\theta_s} \right)^{2b+3} \tag{6.5a}$$

with one scaling parameter b, and also that by Van Genuchten *et al.* (1980):

$$K = K_s s^l \left(1 - \left(1 - s^{1/m} \right)^m \right)^2 \tag{6.5b}$$

with $s = \dfrac{\theta - \theta_r}{\theta_s - \theta_r}$, which has more parameters (l, m) and thus increases the flexibility in the form of the conductivity curves. The moisture-dependence of K expresses the change of the suction forces acting on the soil water exerted by the soil matrix (also denoted as the *hydraulic head*). Similarly, expressions of the hydraulic diffusivity D_w are expressed using a value at saturation and one or more shape parameters.

In these equations some scaling values of the soil water content are used; these parameters are important in defining the effective dynamic range of soil water, and thus in defining the effective water storage capacity of the soil. The following key variables are generally found in land surface schemes, although their naming and definition may vary between different applications:

(a) The *saturation soil moisture content* θ_s is roughly similar to the *soil porosity*, or the relative volume of a soil parcel that can be filled with water. It varies between 0.2 for sandy soils to 0.5 for loam.

(b) The *field capacity* is used to denote the water content that is reached when the free gravity drainage of a saturated soil has ceased, normally two to three days after saturation. It represents the water content value when an equilibrium exists between the gravity force and the Van der Waals forces (adhesion of water to solids). Plant scientists tend to define the field capacity as the point at which plants start to respond to a shortage of soil moisture, which is not necessarily the same value as the soil definition. In many land surface schemes, however, the field capacity is a single value used to control both the canopy stress and the soil hydraulic functions. Its value varies between 0.2 and 0.35.

(c) The (*permanent*) *wilting point* is defined in relation to the behaviour of plants: it is the soil water content at which plants are no longer able to extract water from the soil. Typically it is around or lower than 0.1

(d) The *residual soil moisture content* θ_r is an amount of water that can not be extracted from the soil. It is used in some forms of the Van Genuchten set of equations as another shape parameter in the conductivity/diffusivity curves.

(e) The *wetness index* or *moisture index* is a relative stress indicator. It is a normalized soil moisture content between two thresholds, normally the field capacity and wilting point.

The *dynamic range* of soil moisture is largely determined by the settings of these scaling variables. At water contents greater than field capacity, additional soil water is rapidly removed by free gravity drainage, whereas below wilting point vertical motion of soil water virtually stops.

6.3.4 Calculation of the energy fluxes

Prior to the calculation of the tendencies of the soil water content, it is necessary to specify the rates at which water is extracted by canopy transpiration or bare ground evaporation. The calculation of these terms is linked to the solution of the surface energy balance equation (equation (6.1)), which forms the heart of many land surface parameterization schemes. There are many numerical ways to solve the surface energy balance equation. A simple procedure is to consider the net available energy ($R_n - G$) as given, and to use the Penman-Monteith equation (section 3) to solve for λE and H. However, numerical stability and energy conservation require somewhat more sophisticated schemes that solve for a surface temperature T_{sk} which defines the surface fluxes such that it closes the surface energy balance. As an illustration, we follow the procedure applied in the Tiled ECMWF Surface Scheme for Land (TESSEL; Viterbo & Beljaars, 1995; Van den Hurk *et al.*, 2000), which for special cases comes down to a solution of the Penman-Monteith equation.

In TESSEL, a skin layer with zero heat capacity is used as the interface between the soil and the overlying atmosphere. All terms in equation (6.1) depend in some way on the surface skin temperature T_{sk}. The net radiation is written as:

$$R_n = K\downarrow - K\uparrow + L\downarrow - \varepsilon\sigma T_{sk}^4 \tag{6.6}$$

whereas the soil heat flux reads:

$$G = \Lambda_{sk}(T_{sk} - T_{s1}) \tag{6.7}$$

where Λ_{sk} is a soil heat conductance (units W m^{-2} K^{-1}) and T_{s1} is the temperature of the top soil layer.

The latent heat flux is governed by a moisture gradient between the surface and the lowest layer in the atmospheric model (q_a), and depends on a set of exchange coefficients expressing the aerodynamic turbulent transfer (C_H), wind speed U and surface dependent regulation of evaporation (α_a and α_s), either via a regulation of the canopy transpiration, a limitation of bare ground evaporation by lack of available soil moisture, or other evaporation constraints:

$$\lambda E = \lambda\rho C_H U\left(\alpha_a q - \alpha_s q_s(T_{sk})\right) \tag{6.8a}$$

The quantity $1/C_H U$ is also expressed as an *aerodynamic resistance for heat and water* *exchange*, $r_{a,H}$. The values of coefficients α_a and α_s are 1 for potential evaporation (interception reservoir, sublimation from exposed snow), and a function of the *surface resistance* for other surface types:

$$\alpha_s = \alpha_a = \frac{r_{a,H}}{r_{a,H} + r_s} \tag{6.8b}$$

with r_s the surface resistance. In earlier versions of TESSEL (Viterbo & Beljaars, 1995) bare ground evaporation was controlled by expressing α_s as function of the soil moisture content and setting α_a = 1. Later this was changed into a resistance formulation similar to equation (6.8b), with r_s dependent on soil moisture in the top layer.

Finally, H is computed with a similar gradient–exchange coefficient formula:

$$H = \rho C_H U \left(C_p T + gz - C_p T_{sk} \right) \tag{6.9}$$

where the quantity $C_pT_a + gz$ is known as the *dry static energy* of the atmosphere and C_p the heat capacity of air.

For a given set of exchange coefficients, Λ_{sk}, $r_{a,H}$ and r_c, one can solve equation (6.1) by substituting the flux terms in equations (6.6) to (6.9), linearize T_{sk}^4 and $q_s(T_{sk})$ around T_{sk}, and re-arrange in order to bring all terms with T_{sk} to one side of the equation.

An elementary difference between land surface schemes is how the various fractions of the evaporation (transpiration, evaporation of intercepted or soil water, and snow evaporation) are addressed. In some LSMs the definition of the exchange coefficients uses a weighting procedure in which contributions from these different fractions are all captured in a single effective value of the exchange coefficient, and a single energy balance equation is solved for the entire grid box. In other schemes, like TESSEL, the grid box is separated into various sub-fractions (or tiles) with different values of these coefficients, and for each fraction a separate energy balance is calculated. Figure 6.5 shows a schematic of this approach. The following approaches are followed for each tile.

In TESSEL, two tiles are used to represent *vegetation*: one for the forest area, one for the low vegetation area. The surface resistance r_s is similar to a canopy resistance r_c which is an empirical function:

$$r_c = \frac{r_{s,\min}}{LAI} f_1 f_2 f_3 \tag{6.10a}$$

where $r_{s,\min}$ is a minimum stomatal resistance depending on the specified vegetation type present in the grid box fraction (see next section), *LAI* the Leaf Area Index (expressing the potential amount of evaporating area per unit ground area), and the *f*-functions empirical functions of (photosynthetically active) radiation, (liquid) soil water content W (weighted over the multiple soil layers using a predefined rooting profile), and atmospheric humidity. This approach is often referred to as the *Jarvis-Stewart* approach, being authors that worked extensively on these functions in the late 1970s and 1980s.

An alternative procedure to calculate r_c is to use a sub-model which computes the photosynthetic rate A, and a CO_2 concentration gradient between the air within the stomata and the ambient air Δc, from which r_c can be derived as:

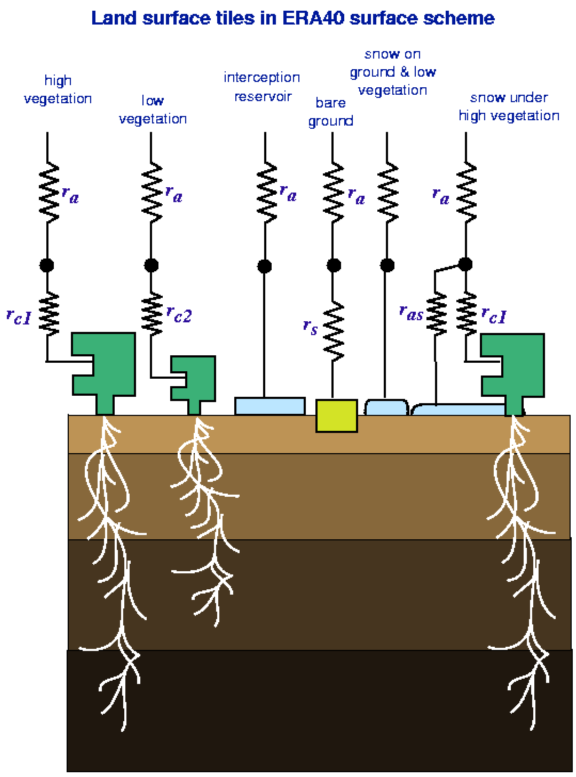

Fig. 6.5 Schematic layout of an example land surface model, in this case the Tiled ECMWF Scheme for Surface Exchange over Land (TESSEL; Van den Hurk et al, 2000).

$$r_c = \frac{M_w}{M_c}\frac{\Delta c}{A} \tag{6.10b}$$

M_w/M_c is the molecular weight ratio between water and CO_2. This procedure thus couples the fluxes of CO_2 (via A) and water (via the transpiration rate), and gains interest in land surface modelling because it includes a response of vegetation to changed ambient CO_2 concentrations. Sophisticated hybrid formulations fully coupling water and CO_2 transport exist as well, where r_c is still expressed as function of atmospheric moisture deficit, and soil water following a Jarvis-Stewart approach, and where A is governed by light, temperature and maximum CO_2 transport through r_c (Daly *et al*., 2004)

The *bare ground evaporation* is treated similarly, albeit that r_s is now only dependent on the soil water content in the top soil layer.

Evaporation from the *interception and exposed snow reservoirs* are treated as potential evaporation, that is, by setting r_s to zero. The evaporation rate is then controlled by the aerodynamic exchange process and a humidity gradient between the surface and the atmosphere. In the snow fraction, the net shortwave radiation flux is also altered, by adopting a (dynamic) snow albedo.

In TESSEL a special treatment is given to grid box fractions where *forest is covered with snow*. It represents the situation where the snow is actually situated *below* the canopy space, and where the major part of net radiation is absorbed by the forest canopy which is given a snow-free albedo value. Thus, in this tile, two parallel water fluxes are calculated: one from the forest canopy (which is treated similarly to the high vegetation tiles above), and one from the underlying snow. This secondary evaporation term affects the air humidity in the canopy space, and may thus affect the transpiration rate from this vegetation component. This extra node (see Fig. 6.5) makes the transpiration rate from this forest fraction different from the snow-free forest tile.

After calculating the energy balance equation for each tile, an area weighted flux is defined which is used to drive the atmospheric evolution.

6.3.5 Determination of the exchange coefficients for water and heat in the atmosphere

A brief outline of the calculation of r_s has already been given in the previous section. There is a wealth of literature on the functional shape, calibration and parameter assignment of this surface resistance term, often guided by *in situ* measurement campaigns. The aerodynamic exchange coefficient $C_H = 1/Ur_{a,H}$ expresses the efficiency of the transport of heat or moisture away from or towards the surface by turbulent mixing. In most LSMs it is based on the so-called Monin-Obukhov similarity theory, which is used to relate a turbulent flux (of heat, moisture, momentum) to a vertical gradient (of temperature, specific humidity, wind speed). Turbulent exchange is enhanced: (a) with increasing wind speed and/or surface roughness, i.e. when the roughness of obstacles at the surface cause a drag on the flow, and (b) when there is vertical motion due to buoyancy, where warmer (or moister) and thus lighter air parcels near the surface are moving upward. Therefore, without giving the full derivation of the equations, C_H depends on the following features (see also Chapters 2 and 3):

(a) The *momentum roughness length* z_{0M} of the surface: the higher this value, the rougher the terrain, the higher C_H.
(b) The *thermal roughness length* z_{0H}, which has an equivalent meaning but for heat transport. z_{0M} and z_{0H} are not necessarily equal: over many land surfaces z_{0H} is an order of magnitude lower than z_{0M}. Consider a sparse canopy site with scattered plant obstacles in a flat bare ground field. The obstacles' heights will be an important determinant of z_{0M} as they will affect the mean height of the momentum sink. However, the typical heat source is located at the bare ground in between the plant obstacles, and the mean height of this source may be significantly lower than the height of momentum sink. Again, the higher z_{0H}, the higher C_H.
(c) The (virtual potential) *temperature gradient*, which determines the contribution of buoyancy to the turbulence near the surface. This dependence complicates the solution of the surface energy balance somewhat, since formally an iteration is needed to solve simultaneously for the temperature gradient between the surface and the atmosphere, the exchange coefficient and the heat flux. Many schemes use the temperature gradient from the previous time step to define C_H.

6.3.6 Specification of the surface characteristics and surface tiling

Many surface characteristics vary in time, and must be updated in every time step (for instance: seasonal changes of the presence of vegetation, albedo, roughness length;

dynamic evolution of for instance the presence of snow or intercepted water). These characteristics are determined prior to the time-dependent calculations,.

During this procedure the land surface scheme is prepared to deal with various *subgrid fractions of surface types*, or *tiles*. Land surfaces are highly heterogeneous at the spatial scale of climate models, and various fractions of the grid-box area may have very different characteristics. For instance, light snowfall does not necessarily cover the whole grid-box at once, but fractions of snow-covered and snow-free surfaces may co-exist within a single model grid-box. The same is true for intercepted precipitation, or the simultaneous presence of bare ground and vegetation.

So, the following steps must be followed to assign the various tile fractions:

(a) Define the *a priori* fraction of bare ground and vegetation. This may either be a static field provided from external sources, or calculated as function of time by other components of the climate model (see below). The static fields are usually derived from remotely-sensed data products providing high resolution land-use maps.

(b) Calculate the fraction of snow covered area using the snow depth and a critical value at which the whole grid box fraction is covered with snow (~15 mm). Assign fractions of snow and snow/forest using the *a priori* vegetation fractions.

(c) Repeat this exercise for the interception reservoir, where the maximum interception depth is defined by the LAI.

(d) Assign the remainder of the grid-box not covered with snow or interception to fractions of low/high vegetation and bare ground, using the original *a priori* ratio of these.

(e) Assign vegetation types (and corresponding values like $r_{s,\min}$) to these fractions.

6.4 CALIBRATION, VALIDATION AND APPLICATION IN CLIMATE MODELS

As becomes evident from this chapter, land surface models contain many parameters that need calibration. Some parameters may seem universal (such as the stability functions in aerodynamic exchange coefficients, or snow parameters), others may be very site specific (soil hydraulic quantities) or dependent on plant species (albedo, canopy resistance parameters). A general strategy is to use local field experiments to calibrate all these parameters for specific land surface schemes and surface types. When the parameters are expected to vary spatially, ancillary maps of soil texture or vegetation type are used to distribute these calibration parameters over the areas where the models are applied.

However, occasionally a “blind” application of calibration parameters to locations for which they are not directly derived introduces the need to apply additional corrections to these parameters to improve the model skill. Two examples are given here, to illustrate the difficulty with calibration.

Viterbo *et al*. (1999) modified the dependence of the aerodynamic exchange on atmospheric stability in their global implementation of the land surface scheme in a numerical weather prediction (NWP) model. In particular, the exchange coefficient was increased drastically for stable conditions, compared to the value of the coefficients which were derived from a careful analysis of flux–gradient relationships using observations from a long-term monitoring programme at Cabauw, in The Netherlands.

The coefficients finally implemented operationally generated significantly more vertical exchange at a given value of the Richardson number (a stability index) than was observed at Cabauw. Why was this modification necessary? The problem here is that the flux–gradient parameterization can relatively easily trigger a positive feedback loop, in which a stably-stratified atmosphere reduces downward heat transport to the surface, which increases the stability and further reduces the heat exchange, until a radiative equilibrium is reached and the surface and atmosphere are greatly uncoupled. This feedback loop *sometimes* occurs in the real world, but not as often as in the original implementation of the model. Apparently, the model misses some heat exchange processes (e.g. intermittent turbulence, gravity wave drag, radiative exchange) that reduce the likelihood of an uncoupled surface. The artificial boost of the exchange coefficient mimics these additional processes sufficiently without explicitly representing (or even understanding) them, but introduces a gap between the measured and implemented calibration parameters.

Another example is the calibration of the dynamic range of soil water by means of a calibrated value of the wilting point and field capacity. Calvet *et al*. (2004) found that a modification of the wilting point was needed to simulate the observed annual cycle of soil water content. The relation between a soil texture class and the hydraulic coefficients (equation (6.5)) obtained using measurements normally shows a rather large range of curves that are averaged. The width of the range indicates that there are more factors than texture that determine the hydraulic head of water (e.g. hysteresis, inhomogeneous soil samples, worm holes, vertical stratification), and this is likely to also be true when adopting a texture-class-dependent curve at a local site. Moreover, the dynamic range of soil water is not just a function of the storage capacity in the soil: it also depends on the variability of $P - E$ on a monthly or seasonal timescale. For instance, a soil with a given field capacity and wilting point usually has a very efficient vertical exchange of moisture when the hydroclimate causes it to be constantly fairly wet, while the same soil exhibits very poor drainage characteristics in a dry climate. The final effective control of the soil on the evaporation rate thus also depends on the environmental condition in which it is located, and this fact makes calibration and the transfer of coefficients untrivial.

6.5 INTERFACES WITH OTHER COMPONENTS IN CLIMATE MODELS

Land surface processes are essential components in meteorological weather and climate models. They are tightly linked to the lower atmosphere via the *Planetary Boundary Layer* (PBL). In addition, many climate models complement the meteorological calculations with representations of the evolution of vegetation biomass or even the competition between vegetation species, or modules tracking the runoff via river routing schemes. These links are discussed briefly in the following.

6.5.1 Coupling to the PBL

The PBL is the turbulent atmospheric layer that directly senses the influence of the land surface with its diurnal cycle of the surface energy budget, roughness variations, etc. Typically, it extends to between 50 and 2000 m in height. Vertical mixing is enhanced via turbulence generated by the shear stress due to friction with the surface obstacles, and by buoyancy caused by heating of the air from below. In meteorological models this vertical mixing is usually represented as a (turbulent) diffusion process, in which mixing occurs over a vertical gradient with a turbulent diffusivity *K*. For this, a vertical grid is defined where the prognostic quantities (temperature, wind speed, humidity, …) are calculated at the central grid nodes (or *full levels*), and the fluxes halfway between these nodes (at the *half levels*). The surface is considered to be the lowest half level, and the fluxes at the surface provide the lower boundary condition to the vertical diffusion scheme in the PBL parameterization.

The interaction between the surface and the PBL is thus two-way: the atmospheric state at the lowest full level provides the reference conditions to the surface (the *dynamic forcing*), and conversely the surface fluxes determine the state of the atmosphere. This two-way coupling is normally solved in sequential time steps, but numerical stability arguments require a careful consideration of the way the surface and the PBL scheme exchange information. This is not detailed here.

A physical implication of the two-way coupling between the surface and the atmosphere is that the behaviour of land surface schemes may depend on the overlying PBL scheme or the way they are coupled. Evaporation humidifies the atmosphere, and this provides a negative *feedback* to the evaporation rate since the moisture gradient between the surface and the atmosphere is reduced. The rate of humidification, however, depends not only on the evaporation rate, but also on the efficiency of mixing away this water vapour via the PBL. Thus, land–atmosphere feedback is relatively weak (“poorly coupled”) when efficient mixing results in an atmosphere that seems to act as an infinite sink of water vapour. The reverse is true when mixing is limited due to a strong capping inversion at the top of the PBL. Since the early 1980s the degree of land–atmosphere coupling has been investigated and expressed by means of a *coupling coefficient* Ω (McNaughton & Spriggs, 1986; Jacobs & de Bruin, 1995; Ek & Holtslag, 2004; Koster *et al*., 2004). This coupling coefficient may have a strong impact on the way land surface schemes behave in the coupled meteorological models. This leads to concerns, since the present strategy of calibrating land surface models by means of offline experiments (see section 6.4; “offline” means a land surface model decoupled from an atmospheric model: atmospheric forcings are prescribed and are not affected by the land surface model) will possibly lead to an erroneous representation of the sensitivity of land surface fluxes to environmental conditions.

6.5.2 Vegetation biomass evolution

Vegetation biomass grows and decays. A clear seasonal cycle is associated with the growth and decay of above-ground leaf area, which affects the surface albedo and evaporation characteristics. But, LAI may also vary at shorter and longer time scales. Clear interannual variability is observed, correlating with climatological conditions (precipitation and/or available radiation). And, at shorter time scales, catastrophic events (frost, flooding, disease) may decrease the green biomass.

The growth of biomass is enabled by the photosynthesis process. Some climate models carry a prognostic evolution of LAI depending on the photosynthetic rate and some decay process. This allows the representation of important (seasonal, interannual) fluctuations of the above-ground biomass, thus explicitly including the interaction between vegetation and land surface exchange processes affecting the atmospheric condition or climate.

Explicit representation of photosynthesis in land surface models also enables study of the direct effect of increased levels of CO_2 on plant evolution and hydroclimate. Plants may respond to increased CO_2 levels by reducing the stomatal aperture in order to avoid water loss, since with larger ambient CO_2 concentrations a similar CO_2 flux can be maintained at lower values of the stomatal conductance. This possibly reduces evaporation, which can cause an extra increase of the surface temperature, thereby increasing the temperature rise in response to increased greenhouse gas emissions. Alternatively, plants will *not* increase the water-use efficiency (assimilated carbon per unit transpired water amount) but respond to increased CO_2 levels by increasing the photosynthesis and thus their growth, thereby increasing evaporation and thus reducing the temperature effect. The vegetation response is likely to be very dependent on ambient climate conditions and the survival strategy of individual plant species.

The principle of coupling environmental conditions → photosynthesis → leaf area index → evaporation → environmental conditions is straightforward, but the practical implementation and the sign of the feedbacks is not. The example of possible vegetation responses to increased CO_2.concentrations illustrates the need to properly represent these processes in climate models and the risk of erroneous settings of parameters that reflect this response, with a strong dependence of the model results on assumptions that are difficult to confirm or falsify in the real world.

6.5.3 River routing

Another example of coupling land surface processes to other relevant processes is given by the explicit modelling of the fate of the runoff water in river systems, aquifers or lakes. Land surface models simply treat the runoff term as an open sink of water, but in reality this water is horizontally transported, eventually into the oceans. For several applications explicit tracking of this water is relevant:

(a) In flood plains, the availability of river water determines the timing and extent of the wetting of the land surface by spilled river water. Short-term flooding events associated with peak-flows in major rivers are simulated in many places in the world since their impact on society may be large. However, seasonal flooding of major swamp areas or wetlands may be important for the climate and hydrological cycle, especially when the flooded area is large and can provide a large source of evaporation water. To simulate these dynamics, a routing scheme is needed that follows the water from the originating source area, via a river/channel system, to the inundation area.

(b) Extensive irrigation may provide another (artificial) source of atmospheric water, not represented by standard land surface modelling. The source of the irrigation water can either be a surface water body (such as the flood plains discussed above), or deep groundwater aquifers. When the irrigation area is large and the amount of irrigation is high compared to the annual precipitation, tracking this water via a horizontal routing scheme is particularly relevant.

6.6 LARGE-SCALE APPLICATIONS: SOME EXAMPLES

In this final section we give two examples of large-scale applications of land surface models: the *Global Soil Wetness Project* (GSWP), and the routine data assimilation of soil water content in the ECMWF (European Centre for Medium-Range Weather Forecasts) model.

6.6.1 The Global Soil Wetness Project

What is the magnitude and dynamic of the terrestrial water balance components at a continental or global scale? Global-scale observations of precipitation can in principle be derived from a combination of remote sensing and (interpolated) gauge measurements, continental-scale runoff can be derived from discharge measurements of major rivers, but continental-scale direct measurements of evaporation or soil water storage are non-existent. Yet the terrestrial water balance is a key component of the global hydrological cycle, and accurate estimation of these terms is highly desirable.

In the Global Soil Wetness Project (see http://grads.iges.org/gswp2/) a wide range of land surface models is used to make an assessment of the temporal and spatial variability of the terrestrial water balance, and to highlight weaknesses in our current understanding of this component and the model uncertainties associated with it. The second edition of GSWP provided forcings of precipitation, radiation, temperature, humidity and wind speed for all global land areas at a spatial resolution of $1° \times 1°$, covering a 10-year period (1987–1996). Approximately 20 land surface models carried out simulations for all these grid points, and a careful analysis of the results is currently taking place. The analyses focus on:

(a) the sensitivity to different sources of the forcing data;
(b) the sensitivity of results to the specification of land surface properties;
(c) inter-model differences;
(d) interannual and spatial variability;
(e) applicability of calibration coefficients at different site locations; and
(f) validation of results with large-scale observational data sets (such as river discharge, snow cover extent, distributed soil moisture and evaporation measurements).

It is well-appreciated that all participating models have deficiencies, and none of them perform optimally in all climate regions or seasons. However, in many validation exercises the mean of the model ensemble seems to provide a fairly robust estimate of the magnitude of the mean and variability of the important terms in the land surface hydrological budget. Globally averaged, the mean depth of the terrestrial storage cycle (snow plus soil water) is approximately 20 cm, while precipitation over land is 840 mm year^{-1}. Approximately 60% of all precipitation is returned via evaporation, the remaining part via runoff. The inter-model differences leave room for ~20% accuracy of these estimates (see Fig. 6.6).

6.6.2 Soil moisture data assimilation

Soil moisture is a slow reservoir in the hydrological cycle, and therefore susceptible to accumulating systematic errors in precipitation, runoff or evaporation. The prognostic evolution of soil moisture content in operational NWP models often results in a clear *drift of the soil water volume*, owing to this accumulation of systematic errors (Viterbo, 1996). Therefore, many operational NWP centres control this drift by repeated corrections of the soil moisture content. Direct observations of soil water are not available at the global scale, thus indirect observations are needed to identify problems with the soil moisture content. At ECMWF and various other NWP centres across the

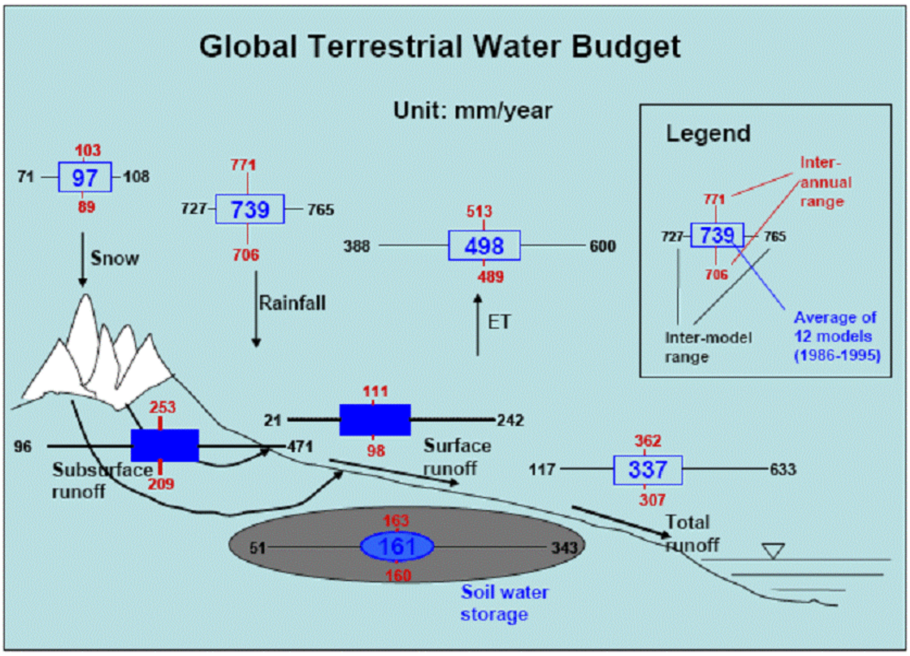

Fig. 6.6 *Mean hydrological fluxes computed from GSWP2 results. Vertical bars indicate the interannual variability, horizontal bars the inter-model variability.*

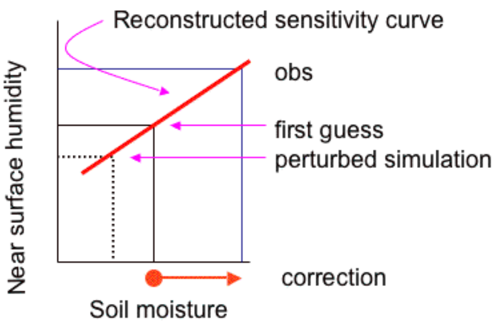

Fig. 6.7 *Schematic of a soil moisture assimilation scheme using observations that are only indirectly related to soil moisture. From a first guess and a perturbed simulation (first guess with a slightly modified soil moisture content) a sensitivity curve between soil moisture and the observation (for instance, relative humidity) is constructed. Via this curve, a new soil moisture is defined that causes an optimal match to the observed soil moisture content.*

world, atmospheric humidity and temperature are used as indirect sources of information on the soil wetness. When there is some coupling between the soil and the atmosphere, a (too) dry soil will result in a (too) low evaporation rate, which gives rise to a (too) high near-surface temperature and (too) low near-surface humidity during daytime. Temperature and humidity are routinely measured, and model errors in these quantities can thus easily be detected (see Fig. 6.7).

The corrected value of soil moisture content in the NWP model is not necessarily comparable to a "true" soil moisture value in the real world. According to the way it is derived in the soil moisture assimilation scheme, it is the model quantity that optimizes the modelled values of temperature and humidity. These quantities are generally well related to soil moisture content, but not always equally, and not necessarily in a manner similar to that in the real world. As such, the resulting (analysed) soil moisture content must be regarded as a *model variable*, highly dependent on the equations defining the evolution of soil moisture and its dependence on the atmospheric state. Errors in these equations, or errors in the near-surface humidity/temperature that are not related to soil moisture content are directly transferred into soil water corrections, and these may be very different from the true wetness state of the land surface.

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