<u>Land-Surface Models</u>

Introduction

A land-surface model must be able to accurately depict the interactions of the atmosphere with the underlying surface land as well as the interactions of the sub-surface, or *substrate*, with the surface. In specific, land-surface models need to accurately predict *heat and moisture transfer* <u>within the substrate</u> as well as <u>between the surface and atmosphere</u>; momentum interaction with the surface is generally viewed as the domain of surface layer models. It must also provide inputs to surface layer and radiation parameterizations to compute surface sensible and latent heat fluxes as well as reflected, absorbed, and emitted shortwave and longwave radiation at the surface.

To be able to accurately predict heat and moisture transfer, however, a land-surface model requires atmospheric inputs. For example, precipitation infiltration can increase soil moisture whereas soil moisture can decrease via evaporation and transpiration to the atmosphere above. In total, a land-surface model requires wind, temperature, precipitation, and radiative forcing input. Consequently, land-surface models are typically coupled to an atmospheric model. There are two such constructs:

- **Direct Coupling**: A land-surface model runs simultaneously with an atmospheric model.
- **Indirect Coupling**: A stand-alone land-surface model integrates surface information with inputs from a stand-alone atmospheric model to develop analyses of relevant land-surface fields (namely, soil temperature and soil moisture).

A land-surface model run in an indirectly coupled fashion is commonly referred to as a land data assimilation system. It is used to assimilate soil state observations and therefore correct initial estimates for soil state fields that are provided by coupled atmospheric-land model forecasts (that we know to intrinsically be imperfect). A conceptual schematic of the relationship between a land data assimilation system, land-surface model, and atmospheric model is provided in Fig. 1.

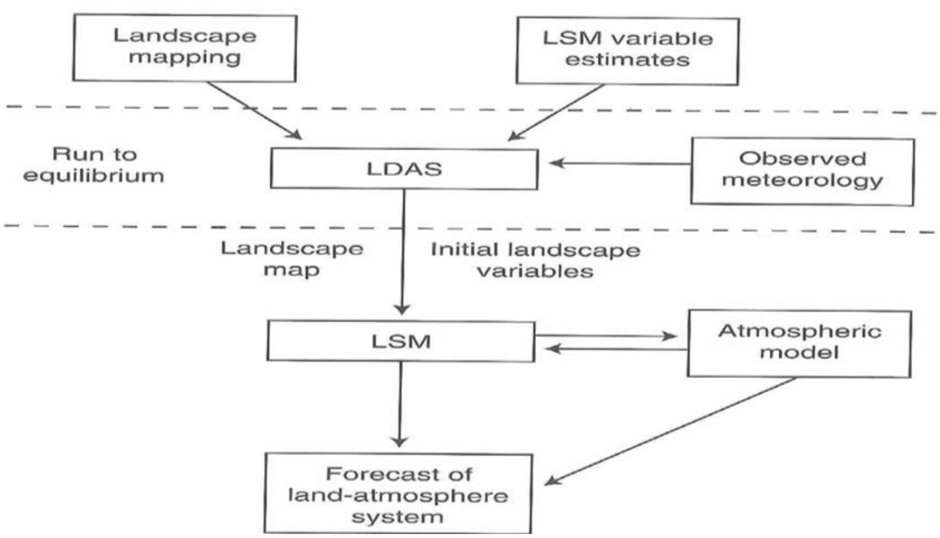

Figure 1. Conceptual schematic of land-surface modeling, including the framework of land data assimilation and its role in providing inputs to a coupled atmosphere-land modeling system. Figure reproduced from Warner (2011), their Fig. 5.6.

Landscape mapping provides information about soil properties, including soil type, vegetation type and cover, terrain height, and other land use characteristics. In general, these are read in from static datasets and are assumed to remain static over the duration of a simulation. However, properties such as green fraction can change on sub-seasonal and longer time scales. For such simulations, it is desirable to be able to update such properties, whether using climatological information or from an external analysis. The static datasets provide information to the land-surface model that it uses, alongside predicted quantities such as soil temperature and moisture, to compute heat and moisture transfer-related coefficients.

Processes Handled by Land-Surface Models

Although heat and moisture transfer in the substrate and at the surface-atmosphere interface may sound straightforward to predict, it is far more complex, with far less known about it and far fewer observations of it, than heat and moisture transfer within the free atmosphere. The major physical processes that a land-surface model must predict in order to accurately predict heat and moisture transfer are depicted in Fig. 2. These can be partitioned into three classes: those occurring in the substrate, at the substrate-surface interface, and immediately above the surface.

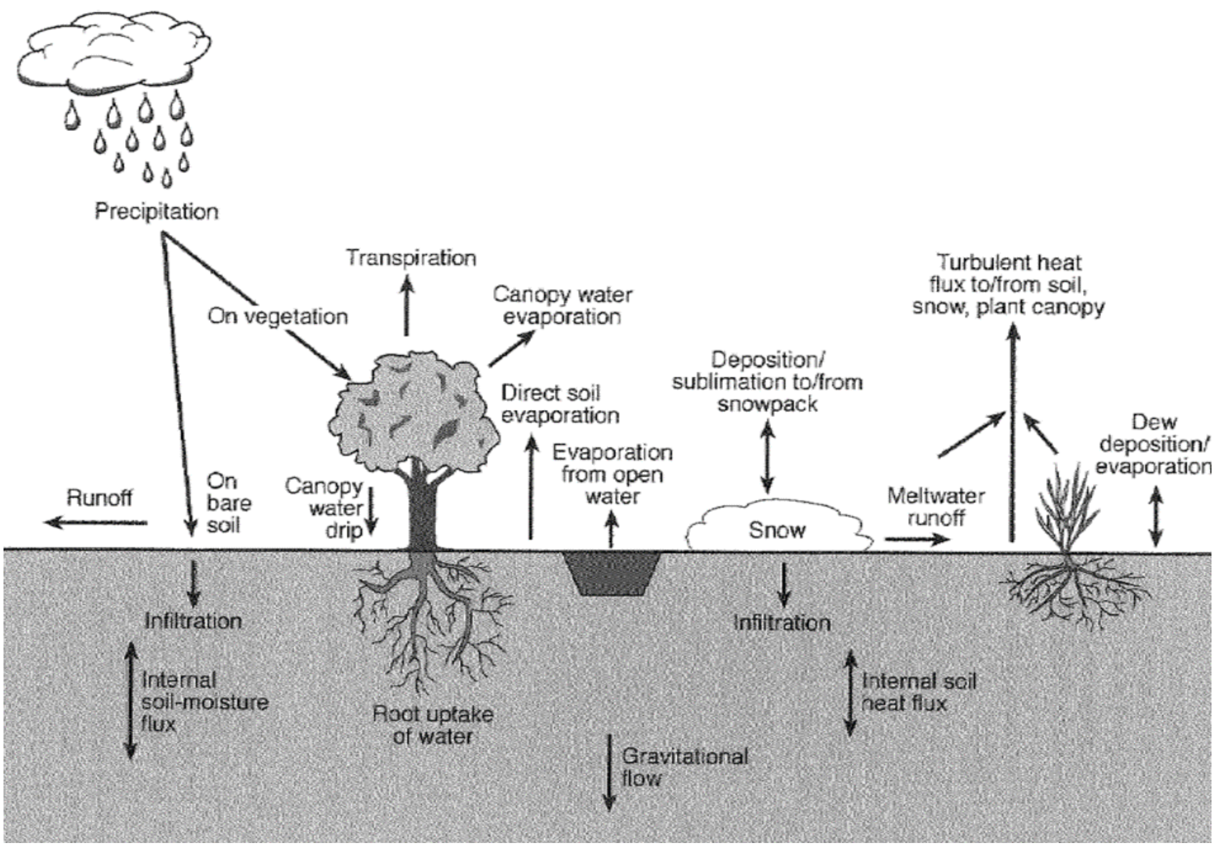

Figure 2. Schematic of the major physical processes that a land-surface model must parameterize in order to accurately predict heat and moisture transfer within the substrate and between the land and the atmosphere. Figure reproduced from Warner (2011), their Fig. 5.1.

In the substrate, freezing, thawing, evaporation, and condensation of substrate water result in the gain or loss of latent heat. Vertical heat transport is primarily accomplished by conduction manifest as a soil heat flux. Vegetation can uptake water, while liquid water can also be transported up or down by changes in water table height, down by gravity, and in all directions by capillary effects. Finally, convection and molecular diffusion can transport water vapor up or down. These processes all are exclusively handled by a land-surface model and only indirectly rely on atmospheric model inputs. Land-surface models may assume that a grid box has uniform or varying soil properties, the choice of which will impact the treatment of these processes within the substrate and just above.

At the substrate-surface interface, surface sensible and latent heat fluxes result in heat exchange between the atmosphere and soil. Moisture content can change as a result of vegetation uptake into stems and leaves, evaporation and sublimation into the atmosphere, and rain, dew, snowmelt, and irrigated water infiltration from above. Runoff and other groundwater flows can also change local moisture content. All except for irrigation require atmospheric inputs but also are modulated by soil properties intrinsic to or predicted by the land-surface model. Irrigation must be parameterized in some fashion, if it is included at all. Runoff and groundwater flows are typically the domain of hydrological models, although land-surface models can parameterize them to varying extents. A land-surface model may include one or multiple layers of vegetation and may or may not include the urban canopy and its unique surface characteristics in this evaluation.

Immediately above the surface, moisture content is a function of precipitation, fog deposition, and water dripping from vegetation above to the surface below. Both moisture and heat are impacted by evaporation, transpiration, snow/frost melting and sublimation, and dew/frost formation. These processes all require atmospheric inputs. The treatment of snow and frost at and immediately above the surface typically varies between models but significantly impacts the surface energy budget.

In the following sections, we seek to discuss the basic physics behind each of these processes. We start at the atmosphere-land interface and work our way down to the substrate. We focus less on variations between land-surface models in how specific terms and parameters are parameterized and more on model fundamentals, namely the equations underlying all parameterizations. A land-surface model will solve for the equations that we will introduce that describe surface and substrate heat and moisture transport, with certain terms estimated or parameterized.

Surface Energy and Moisture Budgets

To first order, surface energy balance, related to surface heating, can be expressed as:

$$R = LE + H + G$$

R is net radiative forcing, LE is net latent heating, H is net sensible heating between the surface and the atmosphere, and G is net sensible heating between the substrate and surface. Land-surface and atmospheric characteristics influence the magnitude and sign of each term. All non-radiative heat transfer at the surface itself is via conduction.

The net radiative forcing can be expressed as:

$$R = (Q+q)(1-\alpha) - I_{\uparrow} + I_{\downarrow}$$

Q is the direct solar radiation incident at the surface and q is the diffuse (or indirect) solar radiation incident at the surface. These are both computed by the shortwave radiation parameterization. α is the albedo, a measure of the surface's reflective characteristics, such that $1 - \alpha$ is the transmittance. This is a function of the underlying surface characteristics, both static and time-varying (e.g., soil moisture, whether or not snow is present, etc.).

$I_{\uparrow}$ is the outgoing longwave radiation flux from the surface and is equal to $\varepsilon\sigma T^4$; thus, it depends on the surface soil temperature. $I_{\downarrow}$ is the absorbed longwave radiation emitted by the atmosphere and is equal to the product of the longwave radiation incident at the surface and the transmittance. Both of these terms are computed by the longwave radiation parameterization.

Though all non-radiative heat transfer at the surface, within the laminar sublayer, is by conduction, vertical transport just above the surface (e.g., within the lowest couple of meters above ground) is primarily accomplished by turbulent vertical eddies driven by buoyancy and vertical wind shear. Both transports occur at rates modulated by their respective *diffusivities*, a measure of the ability for energy to be transported by diffusion. We typically represent both by a single eddy diffusivity even though convection is modulated by molecular and turbulence by eddy diffusivities.

The latent and sensible heat fluxes are given by:

$$H = -\rho c_p K_{Ha} \left.\frac{\partial T}{\partial z}\right|_{ls}$$

$$LE = -\rho c_p K_{Wa} \left.\frac{\partial q}{\partial z}\right|_{ls}$$

Each is a function of the change in temperature or moisture over a finite layer at the surface. This is consistent with the conceptualization of surface transport being by conduction. The subscript of ls on each partial derivative indicates that it is computed in the laminar sublayer. c_p is the specific heat of air at constant pressure, q is specific humidity, and K_{Ha} and K_{Wa} are the eddy diffusivities in air for heat and moisture respectively. Each diffusivity is a function of the atmospheric stability, and thus varies with the meteorology and diurnal cycle, as well as the distance from the surface.

If we assume that the flux magnitudes are approximately constant with height near the surface, these expressions can be rewritten in bulk (i.e., parameterized) form as:

$$H = \rho c_p D_H (T_g - T_a)$$

$$LE = \rho l_v D_w \left(q_{s(T_g)} - q_a\right)$$

l_v is the latent heat of vaporization (for unfrozen soils), $q_{s(Tg)}$ is the saturation specific humidity of the soil, q_a is the atmospheric specific humidity, T_g is the surface soil temperature, and T_a is the atmosphere temperature. D_H and D_W are exchange coefficients for heat and moisture, respectively, and depend on stability, wind speed, and surface roughness. Both fluxes are positive for upward transfer; they have larger magnitudes when the temperature or moisture change between land and atmosphere is largest. In general, they also have larger magnitudes for faster wind speeds and when the atmosphere is more turbulent (e.g., stronger vertical wind shear and lower stability), such that the exchange coefficient magnitudes are larger.

Sensible and latent heat fluxes are often solved by a surface layer rather than land-surface model, although they rely on land-surface model inputs of surface soil moisture, surface soil temperature, and surface use characteristics.

To first order, the surface water budget can be expressed as:

$$\frac{\partial \Theta}{\partial t} = P - ET - RO - D$$

Θ is the dimensionless volumetric soil water content, *P* is water input (i.e., precipitation, snowmelt, deposition, and irrigation), *ET* is evapotranspiration, *RO* is lateral runoff, and *D* is infiltration (or drainage) to the substrate. *P* is obtained from the atmospheric model. Land-surface models seek to parameterize *ET*, *RO*, and *D*, and each typically employ slightly different formulations to do so.

Substrate Heat Transport

Vertical heat transport within the substrate is modulated by the *thermal conductivity*, or the ability of the substance (soil, with composition variations in time and space) to transfer heat, and the *heat capacity*, or the amount of heat required to raise the temperature of a unit volume by 1 K. We refer to the former as k_s and the latter as C_s. Both depend upon soil composition, including both specified (e.g., soil type) and predicted (e.g., soil moisture, soil density) factors. The heat capacity is closely related to the *specific heat* c_s, or the amount of heat required to raise the temperature of a unit mass by 1 K.

Two parameters can be derived from these quantities. The *thermal diffusivity* K_s controls the rate at which a temperature change propagates through a medium, while the *thermal admittance* μ_s is the rate at which a surface can accept or release heat energy.

$$K_s = \frac{k_s}{C_s} \qquad\qquad \mu = \sqrt{k_s C_s}$$

The thermal diffusivity can be viewed as analogous to an exchange coefficient. For the thermal admittance, the admittance of both the soil and atmosphere are important. A higher admittance is associated with a reduced temperature change because heat is transferred efficiently rather than stored locally (where it could affect a temperature change).

Soil moisture imposes a particularly large influence on thermal conductivity and heat capacity and, by extension, thermal diffusivity and thermal admittance, as depicted in Fig. 3. Moister soils have both a greater ability to conduct heat and require more thermal energy to warm by 1 K than drier soils. Thus, both k_s and C_s increase with increasing soil moisture content. Thus results in the thermal admittance also increasing with increasing soil moisture content. However, heat capacity increases more rapidly than thermal conductivity at high soil moisture content, resulting in thermal diffusivity having maximum value at intermediate values of soil moisture.

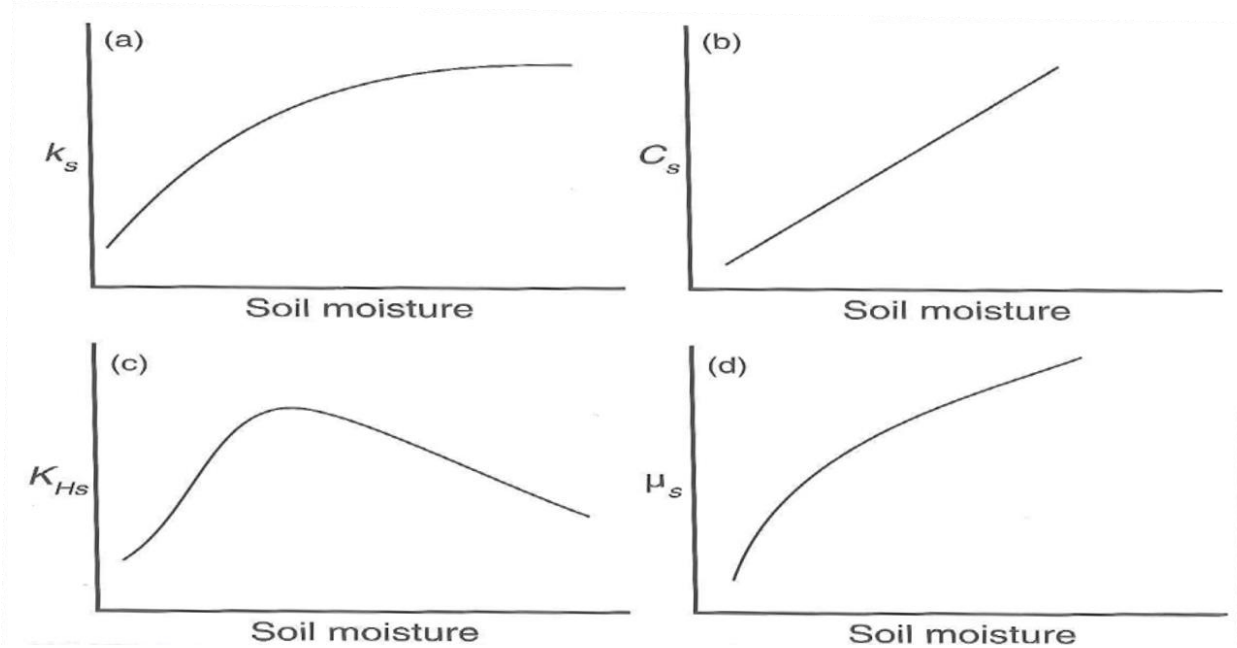

Figure 3. Dependence of (a) thermal conductivity, (b) heat capacity, (c) thermal diffusivity, and (d) thermal admittance on soil moisture. Figure reproduced from Warner (2011), their Fig. 5.2.

Vertical heat transport within the substrate is primarily by conduction, or molecular diffusion, and is down-gradient from high to low values. It can be parameterized like explicit diffusion, wherein the thermal conductivity serves as the diffusion coefficient:

$$H_s = -k_s \frac{\partial T_{substrate}}{\partial z}$$

where H_s is the substrate heat flux. The rate at which substrate temperature changes is related to the vertical convergence of the substrate heat flux and to the heat capacity; i.e., is more heat being fluxed into or out of the layer, and how much of that heat is associated with a 1 K change in the substrate temperature?

$$\frac{\partial T_{substrate}}{\partial t} = -\frac{1}{C_s}\frac{\partial H_s}{\partial z} = \frac{1}{C_s}\frac{\partial}{\partial z}\left(k_s \frac{\partial T_{substrate}}{\partial z}\right)$$

If we make the crude approximation that thermal conductivity is constant with height with in the substrate – i.e., soil composition is constant with height, given its control on thermal conductivity – then we can write:

$$\frac{\partial T_{substrate}}{\partial t} = \frac{k_s}{C_s}\frac{\partial^2 T_{substrate}}{\partial z^2} = K_s \frac{\partial^2 T_{substrate}}{\partial z^2}$$

This relationship indicates that the change in substrate temperature with time is related to its second partial derivative with depth and the soil's thermal diffusivity. We can explore this relationship in idealized forms for day and night, as depicted in Fig. 4.

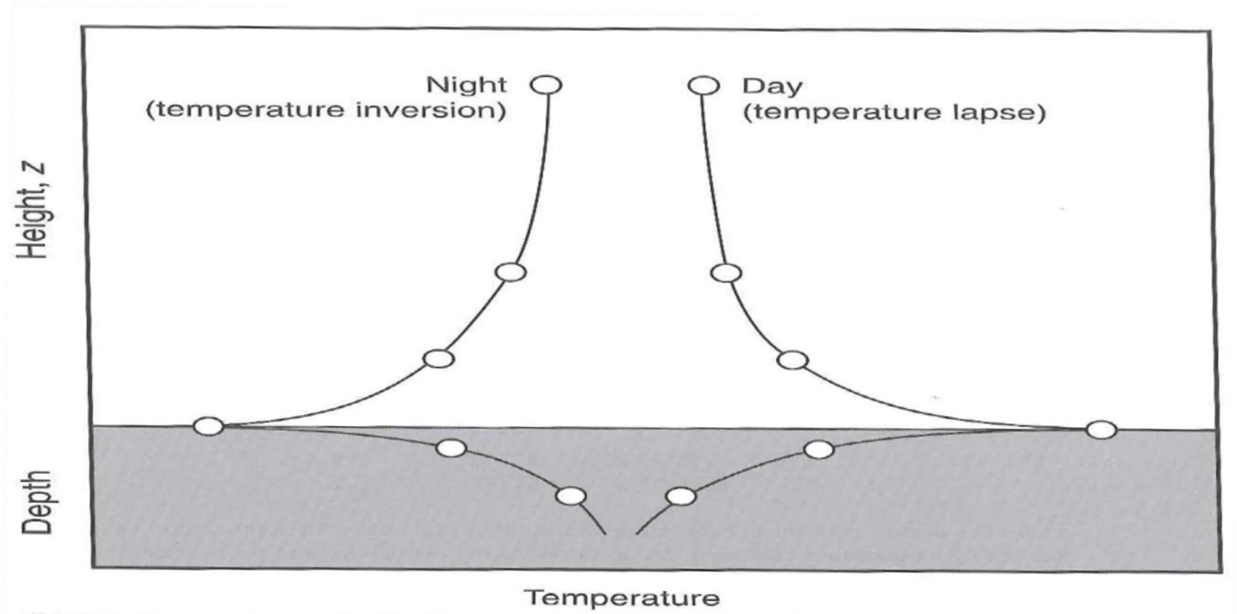

Figure 4. Idealized vertical temperature profiles in the substrate and near the surface at night (left) and during the day (right). These most formally apply to the warm season, when soil temperature typically decreases with increasing depth (averaged over the full diurnal cycle). Figure reproduced from Warner (2011), their Fig. 5.3.

In this idealized scenario, temperature increases toward the surface during the day. Thus, the partial derivative of substrate temperature with height is positive at all depths. However, it is most positive near the surface, where it increases most rapidly with height. This results in the second derivative being positive as well. For positive-definite values of k_s and C_s, this results in substrate warming. Conversely, temperature decreases toward the surface at night, and does so most rapidly near the surface. The partial and second derivatives of substrate temperature are both negative, resulting in substrate cooling. Thus, heat is transported downward during the day to warm the substrate, while heat is transported upward out of the soil at night to cool the substrate.

Substrate Moisture Transport

As noted earlier in these notes, there are six major processes that are relevant to substrate moisture transport. Two apply to water vapor: convection and molecular diffusion. Convection occurs when the substrate temperature lapse rate exceeds the dry adiabatic lapse rate. It permits upward water vapor transport through dry soil by buoyant plumes. Molecular diffusion diffuses water vapor from higher to lower values on the molecular level; it is proportional to the vertical gradient of mixing ratio within the substrate. A land-surface model must parameterize molecular diffusion, given the spatial scales on which it occurs, but may predict convection using an appropriate formulation.

There are four major processes that apply to substrate water transport. The first is associated with changes in water table height below the substrate. A water surface must be in dynamic equilibrium with its surrounding fluid; to achieve this, water will flow laterally from where there is an excess of water to where there is a relative lack of water. This can locally raise the water table where it is initially lower, and vice versa. Fig. 5 provides a schematic of this process for a hypothetical case of thunderstorms inundating the soils along the upslope sides of mountains. At a later time, water is laterally transported beneath the substrate toward the drier valley, raising the water table there while lowering it slightly to the east and west. The rate at which this occurs can be predicted by a land-surface model with accurate representations of soil moisture and water table height as well as formulations for lateral transport rate.

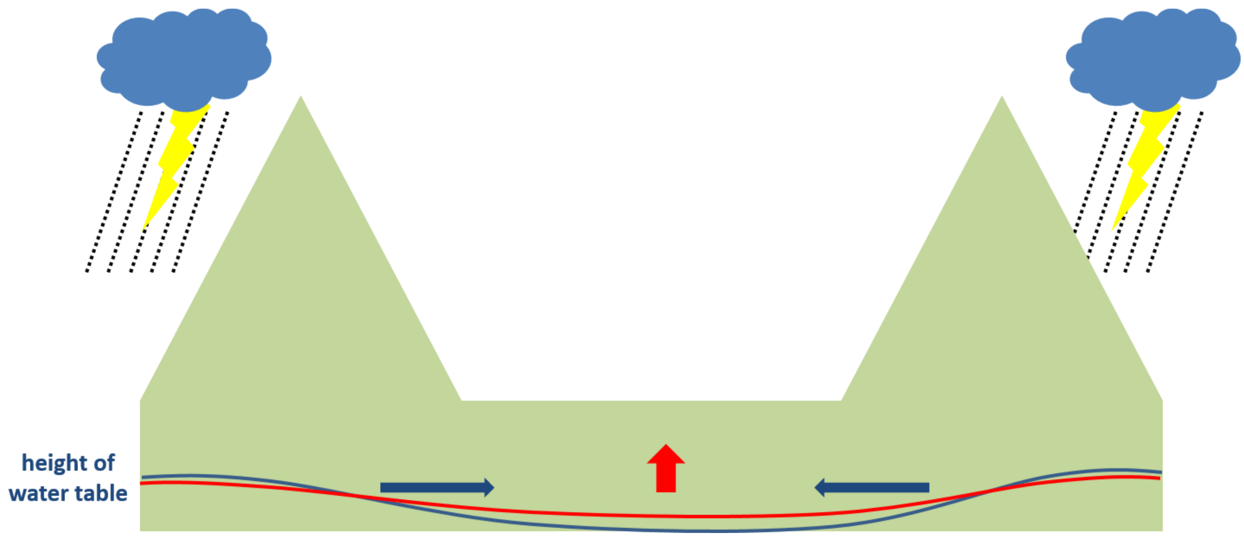

Figure 5. Idealized illustration of thunderstorms contributing to inundation and water table rises (blue) on mountain slopes, after which time lateral transport acts to attempt to restore equilibrium, raising the water table in the valley while lowering it slightly on the mountain slopes (red).

The second is associated with capillary effects, related to surface tension and molecular bonding. Surface tension is a measure of how well water molecules are bonded to soil particles. It is related to the soil's porosity, or how well it can be permeated by water, and to soil moisture content itself. More porous soils have weaker molecular bonding as water molecules can more freely permeate (rather than stick to) the soil. Drier soils also have weaker molecular bonding. Conversely, strong molecular bonding allows water molecules to gradually move vertically by molecular attraction. In general, capillary effects result in movement from moist toward dry soils, particularly between the lower, relatively dry substrate and the water table below it. As with vapor molecular diffusion, these processes must be parameterized by land-surface models.

The third is associated with downward infiltration from the surface and through the substrate. This is driven by gravity, controlled by surface tension, and influenced by the soil moisture potential (a measure of soils' ability to retain water). High surface tension mitigates downward infiltration and low surface tension readily permits downward infiltration. Soil moisture potential is related to soil moisture content and porosity; it is higher for wetter soils and for less porous soils. Vertical water transport occurs more readily for high soil moisture potential; e.g., for saturated, porous soil, soil water more readily flows elsewhere, but will flow less readily for drier, less porous soil. Downward infiltration relies on input from the atmosphere in the forms of rain, snowmelt, irrigation, and dew.

The last is water transport by vegetation, including the extraction of substrate water by plants, the evaporation of surface water collected on the plant canopy, and the transpiration of water vapor out of plant leaves' pores. Though the physics of transpiration are not sufficiently well-understood, we know that it is largely dependent on vegetation type, vegetation density, atmospheric moisture content, the diurnal cycle, and soil temperature and moisture (as relating to plant stresses). All of these processes rely upon accurate surface cover specification and parameterization of the relevant physical processes.

A substrate soil moisture budget can be expressed in terms of a vertical moisture flux, where:

$$\frac{\partial \Theta}{\partial t} = -\frac{\partial q}{\partial z} + E_t$$

Θ is the dimensionless volumetric soil moisture, q is specific humidity, and E_t is a parameterized loss of soil moisture to the plant canopy by evapotranspiration term. The vertical moisture flux can be expressed as:

$$-\frac{\partial q}{\partial z} = \frac{\partial}{\partial z}\left(K_\Theta + D_\Theta \frac{\partial \Theta}{\partial z} \right)$$

K_Θ is the hydraulic conductivity associated with infiltration; it is inversely related to soil moisture potential. D_Θ is the soil water diffusivity associated with surface tension effects. Both are specified by the land-surface model in light of both static and predicted soil characteristics. Different models will use different formulations for these terms; examples given by the course text include:

$$K_\Theta = K_{\Theta s}\left(\frac{\Theta}{\Theta_s} \right)^{2b+3} \qquad D_\Theta = -\left(\frac{b K_{\Theta s} \Psi_s}{\Theta} \right)\left(\frac{\Theta}{\Theta_s} \right)^{b+3}$$

Subscripts of s indicate saturation (i.e., holding the maximum amount of soil moisture) values. b is an empirically derived coefficient. Ψ is the soil moisture potential. Note the strong relationship to soil moisture content and to soil properties, the latter entering through $K_{\Theta s}$ and Ψ_s.

Practical Applications

A land-surface model may employ a surface layer, although many models consider surface layer parameterizations to be separate from land-surface models. In fact, models such as the WRF-ARW model pair surface layer parameterizations with planetary boundary layer parameterizations rather than land-surface models. Land-surface models typically use between one and ten substrate layers between the surface and a specified depth at which the substrate is said to not meaningfully impact the atmosphere above on the time scales of the simulation. For example, the widely used NOAH land-surface model, used by the GFS and NAM models, uses four substrate layers for the 0-10, 10-40, 40-100, and 100-200 cm layers. The RUC land-surface model, used by the RAP and HRRR models, uses nine substrate layers for the 0-1, 1-4, 4-10, 10-30, 30-60, 60-100, 100-160, and 160-300 cm layers. Land-surface models may also use multiple vegetation canopy and/or snow layers.

Each land-surface model has different ways in which it predicts the evolution of soil temperature and soil moisture. This is no different than other parameterized processes. It is well-known that a multiply nested model simulation should maintain parameterization consistency between domains, and this is true for land-surface models as well. However, what about simulations that use another model's output for initial and lateral boundary conditions?

Microphysical parameterizations predict model prognostic variables such as mass mixing ratio and number concentration for various water species. Some planetary boundary layer parameterizations, namely local closure schemes, predict model prognostic variables such as turbulent kinetic energy. Yet, when another model's output is used to initialize or provide lateral boundary conditions for a model simulation, these quantities are typically neglected, and thus it generally does not matter if a different parameterization for these processes is used than was used with the model that provides the initial and lateral boundary conditions.

Cumulus and radiation parameterizations update model prognostic variables such as temperature and water vapor mixing ratio that are routinely observed and can be updated by data assimilation. Though different parameterizations for each set of processes take different approaches to updating model prognostic variables, the updated variables have identical meaning between models. Thus, it generally also does not matter if a different parameterization for these processes is used than was used with the model that provides the initial and lateral boundary conditions.

In contrast, different land-surface models predict soil properties on different levels using different methods and often different representations for the surface and substrate. Consequently, numerical weather prediction model simulations are generally run using the same land-surface model as was used to generate the initial soil state fields. Alternatively, a spin-up period of at least several days may be used to allow the soil state fields to adequately adjust to the particular land-surface model, although the short duration of most weather simulations limits the extent to which this is done.