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Disclaimer

Both the program code and the LISFLOOD documentation (including the LISFLOOD Model Documentation, the LISFLOOD User Guide and the **XXX***) have been carefully inspected before publishing. However, no warranties, either expressed or implied, are made concerning the accuracy, completeness, reliability, usability, performance, or fitness for any particular purpose of the information contained in this documentation, to the software described in this documentation, and to other material supplied in connection therewith. The material is provided "as is". The entire risk as to its quality and performance is with the user.

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About this hydrological model documentation

This Github repository contains the most up-to-date and complete technical documentation of the LISFLOOD model. This includes all it's concepts and model equations of all the standard LISFLOOD processes, but also all the optional modulars. The objective of this model documentation is to provide the interested user with a transparent picture of the hydrological model in order to understand what's behind it.

This document is **not a LISFLOOD user guide!** A LISFLOOD user guide can be found in the lisflood-code repository, together with the code and contains a step-by-step guide on what you need to do and know through the whole chain from the system requirements to receiving the LISFLOOD output. In order to apply this knowledge into practise we have created two use cases, one in Germany (Mulde catchment) and one in Brazil (), which should help you to set up and test the model on your PC. Once you have mastered this step, you can check and verify that the model has been installed and used by you correctly. Then it will be time for you to move onto your own set-up whereever you like to model.

In order to ease the set-up of your own catchment we have created another Github repository called lisflood-utilities that contains all kind of useful tools that should help you preparing your own set-up. Each tool is documented and will explain you how to use it.

Lastly, we also share with you two other tools: 1) LISVAP, our tool to calculate the evapotranspiration and 2) our calibration tool that we are developed. They can be found including user documentation in their respective repositories: **INSERT NAMES HERE**.

Introduction

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The LISFLOOD model is a hydrological rainfall-runoff model that is capable of simulating the hydrological processes that occur in a catchment. LISFLOOD has been developed by the floods group of the Natural Hazards Project of the Joint Research Centre (JRC) of the European Commission. The specific development objective was to produce a tool that can be used in large and trans-national catchments for a variety of applications, including:

- Flood forecasting
- Assessing the effects of land-use change, river regulation measures and climate change

Although a wide variety of existing hydrological models are available that are suitable for *each* of these individual tasks, few *single* models are capable of doing *all* these jobs. Besides this, our objective requires a model that is spatially distributed and, at least to a certain extent, physically-based. Also, the focus of our work is on European catchments. Since several databases exist that contain pan-European information on soils (King *et al.*, 1997; Wösten *et al.*, 1999), land cover (CEC, 1993), topography (Hiederer & de Roo, 2003) and meteorology (Rijks *et al.*, 1998), it would be advantageous to have a model that makes the best possible use of these data. Finally, the wide scope of our objective implies that changes and extensions to the model will be required from time to time. Therefore, it is essential to have a model code that can be easily maintained and modified. LISFLOOD has been specifically developed to satisfy these requirements. The model is designed to be applied across a wide range of spatial and temporal scales. LISFLOOD is grid-based, and applications so far have employed grid cells of as little as 100 metres for medium-sized catchments, to 5000 metres for modelling the whole of Europe and up to 0.1° (around 10 km) for modelling globally. Long-term water balance can be simulated (using a daily time step), as well as individual flood events (using hourly time intervals, or even smaller). The output of a "water balance run" can be used to provide the initial conditions of a "flood run". Although the model's primary output product

is channel discharge, all internal rate and state variables (soil moisture, for example) can be written as output as well. In addition, all output can be written as grids, or time series at user-defined points or areas. The user has complete control over how output is written, thus minimising any waste of disk space or CPU time.

About LISFLOOD

LISFLOOD is a spatially distributed water resources model, developed by the Joint Research Centre (JRC) of the European Commission since 1997. Since then LISFLOOD has been applied to a wide range of water resources applications such as simulating flood prevention and river regulation measures, flood forecasting, drought and soil moisture assessment and forecasting, the impacts of climate and land-use changes on water resources. Its most prominent application is probably within the European Flood Awareness System (EFAS) operated under Copernicus Emergency Management System (EMS).

It's wide applicability is due to its modular structure as well as its temporal and spatial flexibility. The model can be extended with additional modules when need arises, to satisfy the new target objective. In that sense it can be extended to include anything from a better representation of a particular hydrological flow to the implementation of anthropogenic-influenced processes. At the same time the model has been designed to be applied across a wide range of spatial and temporal scales. LISFLOOD is grid-based, and applications so far have employed grid cells of as little as 100 metres for medium-sized catchments, to 5000 metres for modelling the whole of Europe and 0.1° (around 11 km) and 0.5° (around 55 km) for modelling at global scale. The long-term water balance can be simulated (using a daily time step), as well as individual flood events (using hourly time intervals, or even smaller).

Although LISFLOOD's primary output product is river discharge, all internal rate and state variables (soil moisture, for example) can be written as output as well. All output can be written as grids, or time series at user-defined points or areas. The user has complete control over how output is written, thus minimising any waste of disk space or CPU time.

LISFLOOD is implemented in the PCRaster Environmental Modelling language Version 3.0.0 (Wesseling et al., 1996), wrapped in a Python based interface. PCRaster is a raster GIS environment that has its own high-level computer language, which allows the construction of iterative spatio-temporal environmental models. The Python wrapper of LISFLOOD enables the user to control the model inputs and outputs and the selection of the model modules. This approach combines the power, relative simplicity and maintainability of code written in the the PCRaster Environmental Modelling language and the flexibility of Python. LISFLOOD runs on any operating for which Python and PCRaster are available. Currently these include 32-bits Windows (e.g. Windows XP, Vista, 7) and a number of Linux distributions.

Actual infiltration and surface runoff

The actual infiltration INF_{act} $[mm]$ is now calculated as:

$$INF_{act} = min(INF_{pot}, W_{av} - D_{pref,gw})$$

Finally, the surface runoff R_s $[mm]$ is calculated as:

$$R_s = R_d + (1 - f_{dr}) \cdot (W_{av} - D_{pref,gw} - INF_{act})$$

where R_d is the direct runoff (generated in the pixel's 'direct runoff fraction'). If the soil is frozen ($F >$ critical threshold) no infiltration takes place. The amount of moisture in the upper soil layer is updated after the infiltration calculations:

$$w_1 = w_1 + INF_{act}$$

Atmospheric processes and data

Treatment of meteorological input variables

The meteorological conditions provide the driving forces behind the water balance. LISFLOOD uses the following meteorological input variables:

Code	Description	Unit
P	Precipitation	$[\frac{mm}{day}]$
$ET0$	Potential (reference) evapotranspiration rate	$[\frac{mm}{day}]$
$EW0$	Potential evaporation rate from open water surface	$[\frac{mm}{day}]$
$ES0$	Potential evaporation rate from bare soil surface	$[\frac{mm}{day}]$
T_{avg}	Average *daily* temperature	$^{\circ}C$

Note that the model needs *daily* average temperature values, even if the model is run on a smaller time interval (e.g. hourly). This is because the routines for snowmelt and soil freezing are use empirical relations which are based on daily temperature data **Cinzia is that still correct??**. Just as an example, feeding hourly temperature data into the snowmelt routine can result in a gross overestimation of snowmelt. This is because even on a day on which the average temperature is below T_m (no snowmelt), the instantaneous (or hourly) temperature may be higher for a part of the day, leading to unrealistically high simulated snowmelt rates.

Both precipitation and evaporation are internally converted from *intensities* $[\frac{mm}{day}]$ to *quantities per time step* $[mm]$ by multiplying them with the time step, Δt (in *days*). For the sake of consistency, all in- and outgoing fluxes will also be described as *quantities per time step* $[mm]$ in the following, unless stated otherwise. $ET0$, $EW0$ and $ES0$ can be calculated using standard meteorological observations. To this end a dedicated pre-processing application has been developed (LISVAP), which is documented in a separate manual. Insert link to LISVAP manual once it is produced

Rain and snow

If the average temperature is below 1°C, all precipitation is assumed to be snow. A snow correction factor is used to correct for undercatch of snow precipitation. Unlike rain, snow accumulates on the soil surface until it melts. The rate of snowmelt is estimated using a simple degree-day factor method. Degree-day factor type snow melt models usually take the following form (e.g. see WMO, 1986):

$$M = C_m(T_{avg} - T_m)$$

where M is the rate of snowmelt, T_{avg} is the average daily temperature, T_m is some critical temperature and C_m is a degree-day factor $[\frac{mm}{^{\circ}C\ day}]$.

Speers *et al.* (1979) developed an extension of this equation which accounts for accelerated snowmelt that takes place when it is raining (cited in Young, 1985). The equation is supposed to apply when rainfall is greater than 30 mm in 24 hours. Moreover, although the equation is reported to work sufficiently well in forested areas, it is not valid in areas that are above the tree line, where radiation is the main energy source for snowmelt). LISFLOOD uses a variation on the equation of Speers *et al.* The modified equation simply assumes that for each mm of rainfall, the rate of snowmelt increases with 1% (compared to a 'dry' situation). This yields the following equation:

$$M = C_m \cdot C_{Seasonal}(1 + 0.01 \cdot R\Delta t)(T_{avg} - T_m) \cdot \Delta t$$

where M is the snowmelt per time step $[mm]$, R is rainfall (not snow!) intensity $[\frac{mm}{day}]$, and Δt is the time interval $[days]$. T_m has a value of 0 $^{\circ}C$, and C_m is a degree-day factor $[\frac{mm}{^{\circ}C \cdot day}]$.

However, it should be stressed that the value of C_m can actually vary greatly both in space and time (e.g. see Martinec *et al.*, 1998). Therefore, **in practice this parameter is often treated as a calibration constant**. A low value of C_m indicates slow snow melt. $C_{Seasonal}$ is a seasonal variable melt factor which is also used in several other models (e.g. Anderson 2006, Viviroli et al., 2009). There are mainly two reasons to use a seasonally variable melt factor:

- The solar radiation has an effect on the energy balance and varies with the time of the year.
- The albedo of the snow has a seasonal variation, because fresh snow is more common in the mid winter and aged snow in the late winter/spring. This produce an even greater seasonal variation in the amount of net solar radiation

The following Figure shows an example where a mean value of: 3.0 $\frac{mm}{^{\circ}C \cdot day}$ is used. The value of C_m is reduced by 0.5 at 21^{st} December and a 0.5 is added on the 21^{st} June. In between a sinus function is applied

Figure: *Sinus shaped snow melt coefficient (C_m) as a function of days of year.*

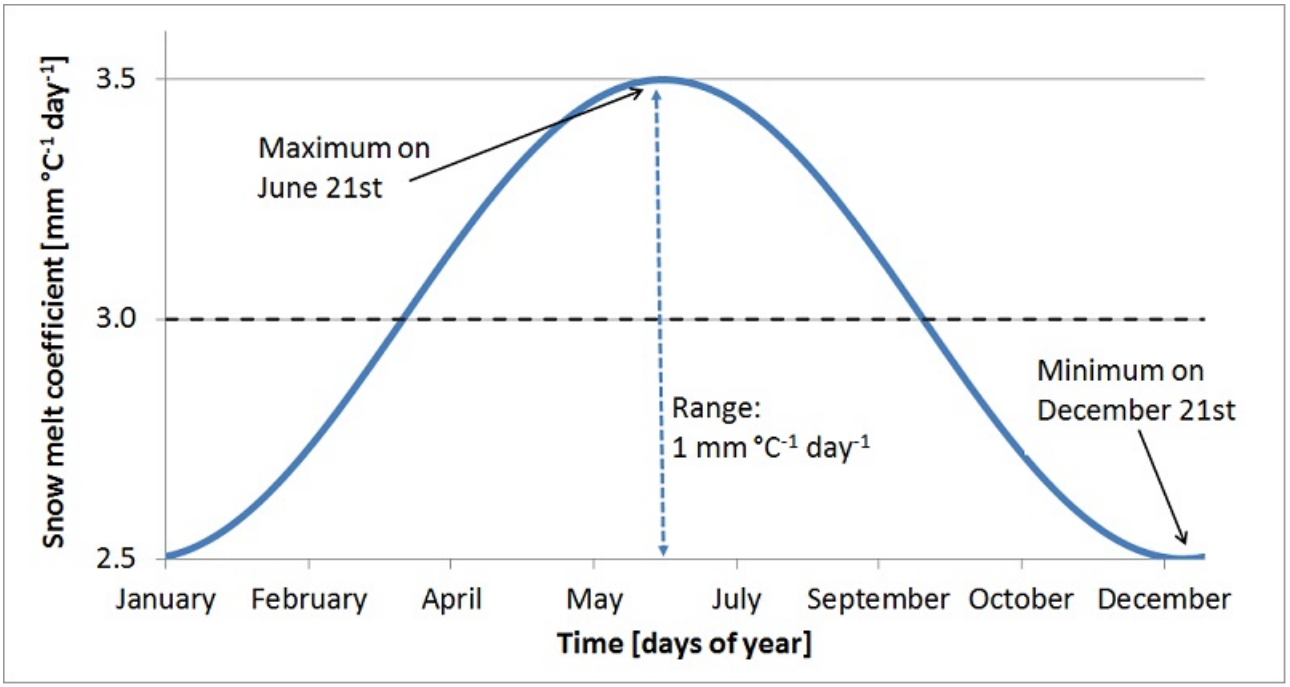

Figure 1:

At high altitudes, where the temperature never exceeds $1°C$, the model accumulates snow without any reduction because of melting loss. In these altitudes runoff from glacier melt is an important part. The snow will accumulate and converted into firn. Then firn is converted to ice and transported to the lower regions. This can take decades or even hundred years. In the ablation area the ice is melted. In LISFLOOD this process is emulated by melting the snow in higher altitudes on an annual basis over summer. A sinus function is used to start ice melting in summer (from 15 June till 15 September) using the temperature of zone B:

Figure: *Sinus shaped ice melt coefficient as a function of days of year.*

The amount of snowmelt and ice melt together can never exceed the actual snow cover that is present on the surface.

For large pixel sizes, there may be considerable sub-pixel heterogeneity in snow accumulation and melt, which is a particular problem if there are large elevation differences within a pixel. Because of this, snow melt and accumulation are modelled separately for 3 separate elevation zones, which are defined at the sub-pixel level. This is shown in Figure below:

Figure: *Definition of sub-pixel elevation zones for snow accumulation and melt modelling. Snowmelt and accumulation calculations in each zone are based on elevation (and derived temperature) in centroid of each zone.*

The division in elevation zones was changed from a uniform distribution in the previous LISFLOOD version to a normal distribution, which fits better to the real distribution of e.g. 100m SRTM DEM pixels in a 5x5km grid cell. Three elevation zones A, B, and C are defined with each zone occupying one third of the pixel surface. Assuming further that T_{avg} is valid for the average pixel elevation, average temperature is extrapolated to the centroids of the lower (A) and upper (C) elevation zones, using a fixed temperature lapse rate, L, of 0.0065 °C per meter elevation difference. Snow, snowmelt and snow accumulation are subsequently modelled separately for each elevation zone, assuming that temperature can be approximated by the temperature at the centroid of each respective zone.

Direct evaporation from the soil surface

The maximum amount of evaporation from the soil surface equals the maximum evaporation from a shaded soil surface, $ES_{max}[mm]$, which is computed as:

$$ES_{max} = ES0 \cdot e^{(\frac{-\kappa_{gb} \cdot LAI}{\Delta t})}$$

where $ES0$ is the potential evaporation rate from bare soil surface $[\frac{mm}{day}]$. The actual evaporation from the soil mainly depends on the amount of soil moisture near the soil surface: evaporation decreases as the topsoil is drying. In the model this is simulated using a reduction factor which is a function of the number of days since the last rain storm (Stroosnijder,

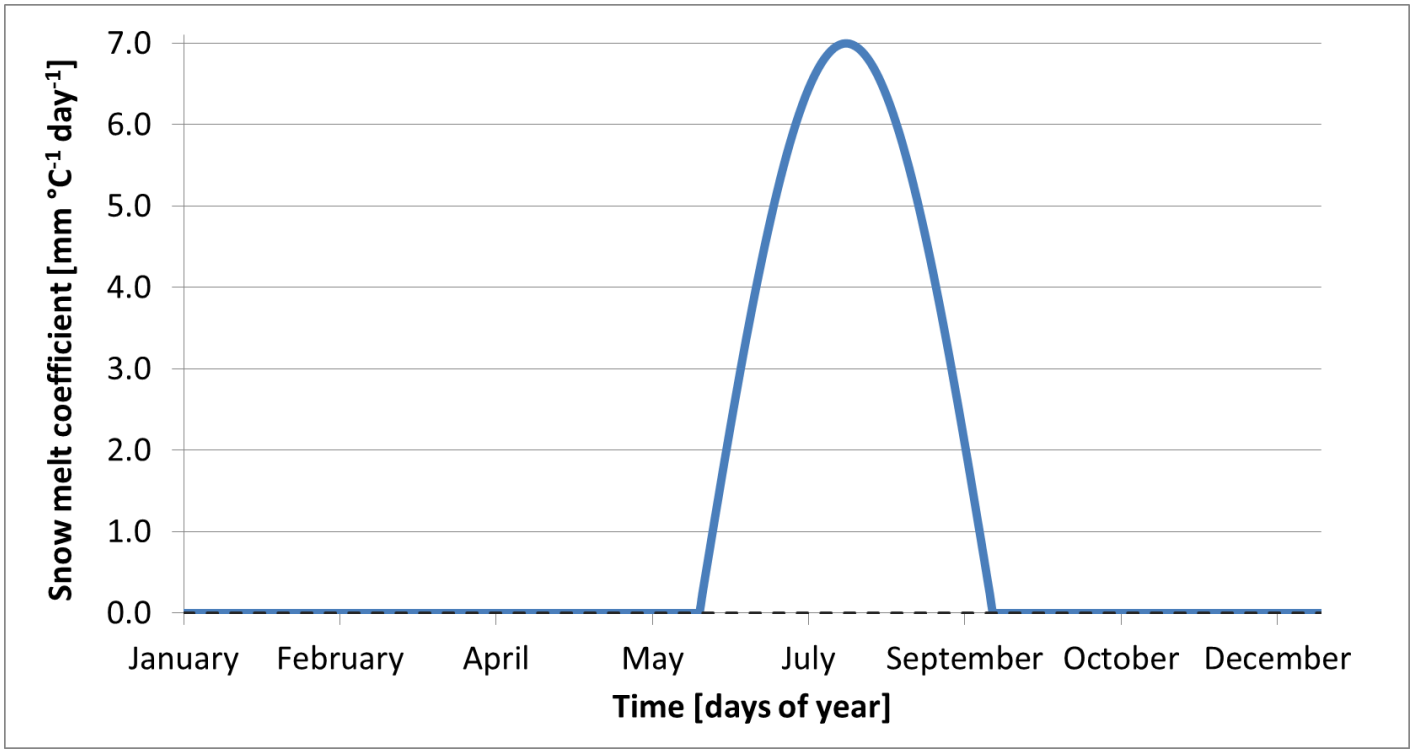

Figure 2:

1987, 1982):

$$ES_a = ES_{max} \cdot (\sqrt{D_{slr}} - \sqrt{D_{slr} - 1})$$

The variable D_{slr} represents the number of days since the last rain event. Its value accumulates over time: if the amount of water that is available for infiltration (W_{av}) remains below a critical threshold it increases by an amount of $\Delta t[days]$ for each time step. It is reset to 1 only if the critical amount of water is exceeded (In the LISFLOOD settings file this critical amount is currently expressed as an *intensity* $[\frac{mm}{day}]$. This is because the equation was originally designed for a daily time step only. Because the current implementation will likely lead to *DSLR* being reset too frequently, the exact formulation may change in future versions (e.g. by keeping track of the accumulated available water of the last 24 hours)).

The actual soil evaporation is always the smallest value out of the result of the equation above and the available amount of moisture in the soil, i.e.:

$$ES_a = \min(ES_a, w_1 - w_{res1})$$

where $w_1[mm]$ is the amount of moisture in the upper soil layer and $w_{res1}[mm]$ is the residual amount of soil moisture. Like transpiration, direct evaporation from the soil is set to zero if the soil is frozen. The amount of moisture in the upper soil layer is updated after the evaporation calculations:

$$w_1 = w_1 - ES_a$$

Evaporation of intercepted water

Evaporation of intercepted water, EW_{int}, occurs at the potential evaporation rate from an open water surface, $EW0$. The *maximum* evaporation per time step is proportional to the fraction of vegetated area in each pixel (Supit *et al.*,1994):

$$EW_{max} = EW0 \cdot [1 - e^{-\kappa_{gb} \cdot LAI}] \cdot \Delta t$$

where $EW0$ is the potential evaporation rate from an open water surface $[\frac{mm}{day}]$, and EW_{max} is in $[mm]$ per time step. Constant κ_{gb} is the extinction coefficient for global solar radiation. Since evaporation is limited by the amount of water stored on the leaves, the actual amount of evaporation from the interception store equals:

$$EW_{int} =_{min} (EW_{max} \cdot \Delta t, Int_{cum})$$

where EW_{int} is the actual evaporation from the interception store $[mm]$ per time step, and $EW0$ is the potential evaporation rate from an open water surface $[\frac{mm}{day}]$ It is assumed that on average all water in the interception store Int_{cum}

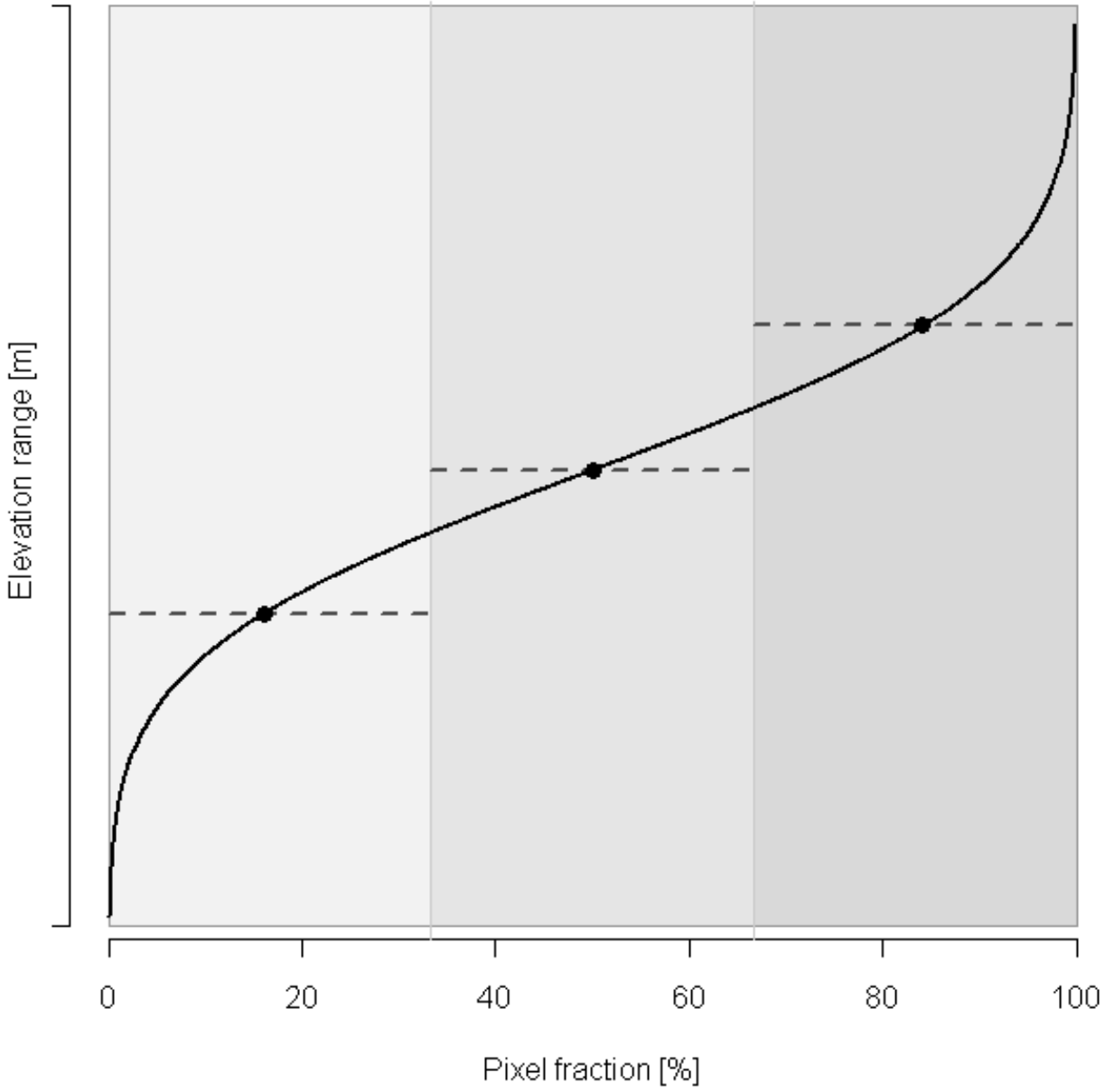

Figure 3:

will have evaporated or fallen to the soil surface as leaf drainage within one day. Leaf drainage is therefore modelled as a linear reservoir with a time constant (or residence time) of one day, i.e:

$$D_{int} = \frac{1}{T_{int}} \cdot Int_{cum} \cdot \Delta t$$

where D_{int} is the amount of leaf drainage per time step $[mm]$ and T_{int} is a time constant for the interception store $[days]$, which is set to 1 day.

Interception

Interception is estimated using the following storage-based equation (Aston, 1978, Merriam, 1960):

$$Int = S_{max} \cdot [1 - e^{\frac{-k \cdot R \cdot \Delta t}{S_{max}}}]$$

where $Int[mm]$ is the interception per time step, $S_{max}[mm]$ is the maximum interception, R is the rainfall intensity $[\frac{mm}{day}]$ and the factor k accounts for the density of the vegetation. S_{max} is calculated using an empirical equation (Von Hoyningen-Huene, 1981):

$$\begin{cases} S_{max} = 0.935 + 0.498 \cdot LAI - 0.00575 \cdot LAI^2 & [LAI > 0.1] \\ S_{max} = 0 & [LAI \leq 0.1] \end{cases}$$

where LAI is the average Leaf Area Index $[\frac{m^2}{m^2}]$ of each model element (pixel). k is estimated as:

$$k = 0.046 \cdot LAI$$

The value of Int can never exceed the interception storage capacity, which is defined as the difference between S_{max} and the accumulated amount of water that is stored as interception, Int_{cum}.

Water uptake by plant roots and transpiration

Water uptake and transpiration by vegetation and direct evaporation from the soil surface are modelled as two separate processes. The approach used here is largely based on Supit *et al.* (1994) and Supit & Van Der Goot (2000). The **maximum transpiration** per time step

$$mm$$

is given by:

$$T_{max} = k_{crop} \cdot ET0 \cdot [1 - e^{(-\kappa_{gb} \cdot LAI)}] \cdot \Delta t - EW_{int}$$

Where $ET0$ is the potential (reference) evapotranspiration rate $[\frac{mm}{day}]$, constant $_{gb}$ is the extinction coefficient for global solar radiation

$$-$$

and k_{crop} is a crop coefficient, a ration between the potential (reference) evapotranspiration rate and the potential evaporation rate of a specific crop. k_{crop} is 1 for most vegetation types, except for some excessively transpiring crops like sugarcane or rice.

> Note that the energy that has been 'consumed' already for the evaporation of intercepted water is simply subtracted here in order to respect the overall energy balance.

The **actual transpiration rate** is reduced when the amount of moisture in the soil is small. In the model, a reduction factor is applied to simulate this effect:

$$r_{WS} = \frac{w_1 - w_{wp1}}{w_{crit1} - w_{wp1}}$$

where w_1 is the amount of moisture in the upper soil layer $[mm]$, $w_{wp1}[mm]$ is the amount of soil moisture at wilting point (pF 4.2) and $w_{crit1}[mm]$ is the amount of moisture below which water uptake is reduced and plants start closing their stomata. The **critical amount of soil moisture** is calculated as:

$$w_{crit1} = (1 - p) \cdot (w_{fc1} - w_{wp1}) + w_{wp1}$$

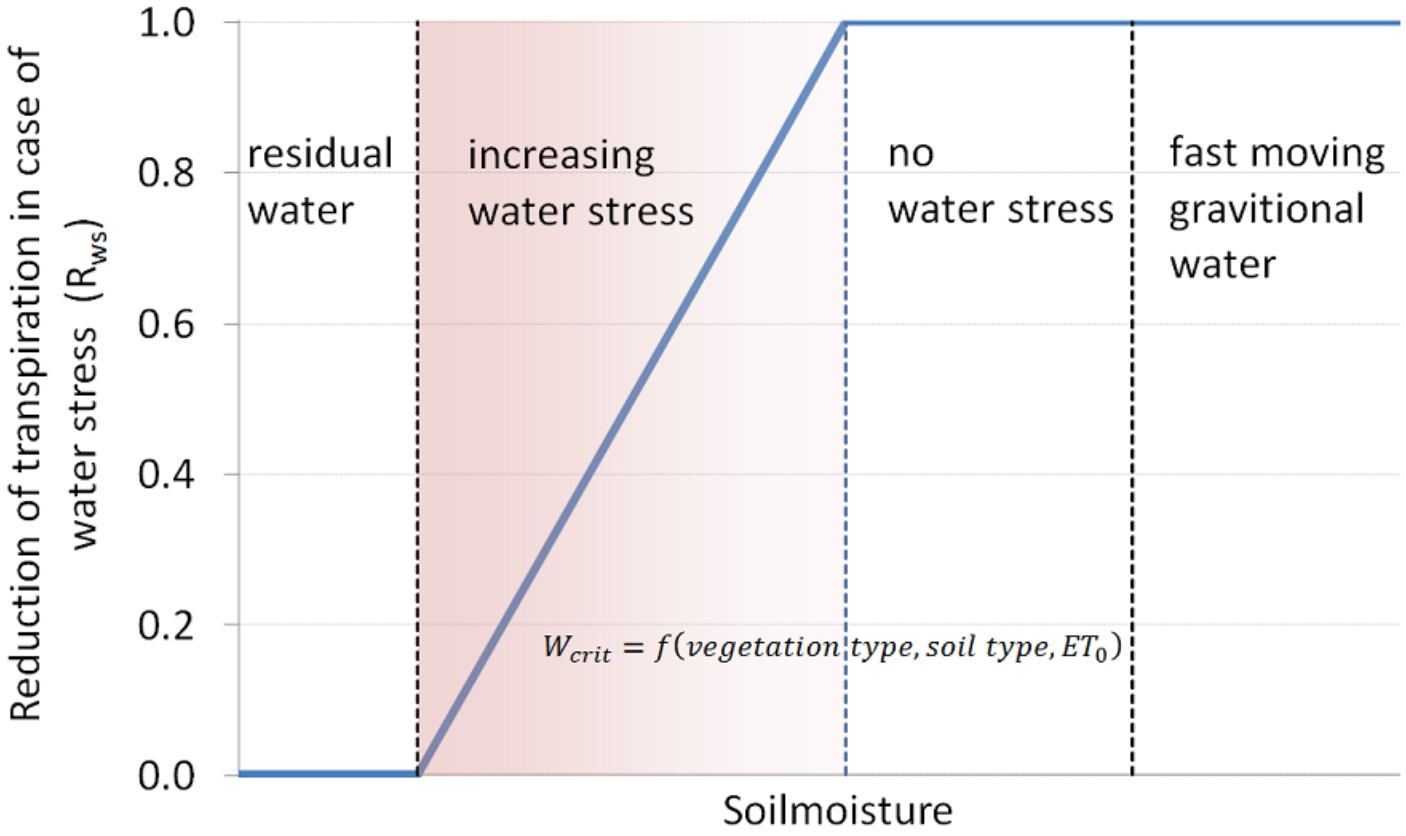

Figure 4:

where $w_{fc1}[mm]$ is the amount of soil moisture at field capacity and p is the soil water depletion fraction. R_{WS} varies between 0 and 1. Negative values and values greater than 1 are truncated to 0 and 1, respectively. p represents the fraction of soil moisture between w_{fc1} and w_{wp1} that can be extracted from the soil without reducing the transpiration rate. Its value is a function of both vegetation type and the potential evapotranspiration rate. The procedure to estimate p is described in detail in Supit & Van Der Goot (2003). The following Figure illustrates the relation between $R_{WS}, w, w_{crit}, w_{fc}, w_{wp}$:

Figure: *Reduction of transpiration in case of water stress. r_{ws} decreases linearly to zero between w_{crit} and w_{wp}.*

The **actual transpiration** T_a is now calculated as:

$$T_a = r_{WS} \cdot T_{max}$$

with T_a and T_{max} in $[mm]$.

Transpiration is set to zero when the soil is frozen (i.e. when frost index F exceeds its critical threshold). The amount of **moisture in the upper soil layer** is updated after the transpiration calculations:

$$w_1 = w_1 - T_a$$

Channel routing

Flow through the channel is simulated using the **kinematic wave equations**. The basic equations and the numerical solution are identical to those used for the surface runoff routing:

$$\frac{\partial Q_{ch}}{\partial x} \cdot \frac{\partial A_{ch}}{\partial t} = q_{ch}$$

where Q_{ch} is the channel discharge $[\frac{m^3}{s}]$, A_{ch} is the cross-sectional area of the flow $[m^2]$ and q_ch is the amount of lateral inflow per unit flow length $[\frac{m^2}{s}]$. The momentum equation then becomes:

$$\rho \cdot gA_{ch} \cdot (S_0 - S_f) = 0$$

where S_0 now equals the gradient of the channel bed, and $S_0 = S_f$. As with the surface runoff, values for parameter $_{k,ch}$ are estimated using Manning's equation:

$$\alpha_{k,ch} = (\frac{n \cdot P_{ch}^{2/3}}{\sqrt{S_0}})^{0.6}; \beta_k = 0.6$$

At present, LISFLOOD uses values for $_{k,ch}$ which are based on a static (reference) channel flow depth (half bankfull) and measured channel dimensions. The term q_{ch} (**sideflow**) now represents the runoff that enters the channel per unit channel length:

$$q_{ch} = \frac{\sum Q_{sr} + \sum Q_{uz} + \sum Q_{lz} + Q_{in} + Q_{res}}{L_{ch}}$$

Here, Q_{sr}, Q_{uz} and Q_{lz} denote the contributions of surface runoff, outflow from the upper zone and outflow from the lower zone, respectively. Q_{in} is the inflow from an external inflow hydrograph; by default its value is 0, unless the 'inflow hydrograph' option is activated. Q_{res} is the water that flows out of a reservoir into the channel; by default its value is 0, unless the 'reservoir' option is activated. $Q_{sr}, Q_{uz}, Q_{lz}, Q_{in}$ and Q_{res} are all expressed in [m^3] per time step. L_{ch} is the channel length $[m]$, which may exceed the pixel size (Δx) in case of meandering channels. The kinematic wave channel routing can be run using a smaller time-step than the over simulation timestep, Δt, if needed.

Evaporation of intercepted water

Evaporation of intercepted water, EW_{int}, occurs at the potential evaporation rate from an open water surface, $EW0$. The *maximum* evaporation per time step is proportional to the fraction of vegetated area in each pixel (Supit *et al.*,1994):

$$EW_{max} = EW0 \cdot [1 - e^{-\kappa_{gb} \cdot LAI}] \cdot \Delta t$$

where $EW0$ is the potential evaporation rate from an open water surface $[\frac{mm}{day}]$, and EW_{max} is in $[mm]$ per time step. Constant κ_{gb} is the extinction coefficient for global solar radiation. Since evaporation is limited by the amount of water stored on the leaves, the actual amount of evaporation from the interception store equals:

$$EW_{int} =_{min} (EW_{max} \cdot \Delta t, Int_{cum})$$

where EW_{int} is the actual evaporation from the interception store $[mm]$ per time step, and $EW0$ is the potential evaporation rate from an open water surface $[\frac{mm}{day}]$ It is assumed that on average all water in the interception store Int_{cum} will have evaporated or fallen to the soil surface as leaf drainage within one day. Leaf drainage is therefore modelled as a linear reservoir with a time constant (or residence time) of one day, i.e:

$$D_{int} = \frac{1}{T_{int}} \cdot Int_{cum} \cdot \Delta t$$

where D_{int} is the amount of leaf drainage per time step $[mm]$ and T_{int} is a time constant for the interception store $[days]$, which is set to 1 day.

Frost index soil

When the soil surface is frozen, this affects the hydrological processes occurring near the soil surface. To estimate whether the soil surface is frozen or not, a frost index F is calculated. The equation is based on Molnau & Bissell (1983, cited in Maidment 1993), and adjusted for variable time steps. The **rate at which the frost index changes** is given by:

$$\frac{dF}{dt} = -(1 - A_f) \cdot F - T_{av} \cdot e^{-0.04 \cdot K \cdot d_s / w \cdot e_s}$$

$\frac{dF}{dt}$ is expressed in $[\frac{^{\circ}C}{day} \cdot \frac{1}{day}]$. A_f is a decay coefficient $[\frac{1}{day}]$, K is a a snow depth reduction coefficient $[\frac{1}{cm}]$, d_s is the (pixel-average) depth of the snow cover (expressed as mm equivalent water depth), and $w \cdot e_s$ is a parameter called snow water equivalent, which is the equivalent water depth water of a snow cover (Maidment, 1993). In LISFLOOD, A_f and K are set to 0.97 and 0.57 $[\frac{1}{cm}]$ respectively, and $w \cdot e_s$ is taken as 0.1, assuming an average snow density of 100 $\frac{kg}{m^3}$ (Maidment, 1993). **The soil is considered frozen when the frost index rises above a critical threshold of 56**. For each time step the value of F $[\frac{^{\circ}C}{day}]$ is updated as:

$$F(t) = F(t-1) + \frac{dF}{dt}\Delta t$$

Note: F is not allowed to become less than 0.

When the frost index rises above a threshold of 56, every soil process is frozen and transpiration, evaporation, infiltration and the outflow to the second soil layer and to upper groundwater layer is set to zero. Any rainfall is bypassing the soil and transformed into surface runoff till the frost index is equal or less than 56.

Groundwater

Groundwater storage and transport are modelled using two parallel linear reservoirs, similar to the approach used in the HBV-96 model (Lindström et al., 1997). The upper zone represents a quick runoff component, which includes fast groundwater and subsurface flow through macro-pores in the soil. The lower zone represents the slow groundwater component that generates the base flow. The **outflow from the upper zone to the channel**, $Q_{uz}[mm]$ equals:

$$Q_{uz} = \frac{1}{T_{uz}} \cdot UZ \cdot \Delta t$$

where T_{uz} is a reservoir constant $[days]$ and UZ is the amount of water that is stored in the upper zone $[mm]$. Similarly, the **outflow from the lower zone** is given by:

$$Q_{lz} = \frac{1}{T_{lz}} \cdot LZ \cdot \Delta t$$

Here, T_{lz} is again a reservoir constant $[days]$, and LZ is the amount of water that is stored in the lower zone $[mm]$. The values of both T_{uz} and T_{lz} are obtained by calibration. The upper zone also provides the inflow into the lower zone. For each time step, a fixed amount of **water percolates from the upper to the lower zone**:

$$D_{uz,lz} = min(GW_{perc} \cdot \Delta t, UZ)$$

Here, GW_{perc} $[\frac{mm}{day}]$ is a user-defined value that can be used as a calibration constant. For many catchments it is quite reasonable to treat the lower groundwater zone as a system with a closed lower boundary (i.e. water is either stored, or added to the channel). However, in some cases the closed boundary assumption makes it impossible to obtain realistic simulations. Because of this, it is **possible to percolate a fixed amount of water out of the lower zone**, as a loss D_{loss}:

$$D_{loss} = min(f_{loss} \cdot \Delta t, LZ)$$

In the previous version of LISFLOOD D_{loss}, was calculated as a fixed fraction of Q_{lz}, but this leads to a high dependency of D_{loss} from GW_{perc} and LZ. For example if either GW_{perc} or LZ is quite low the parameter D_{loss} turns out to be meaningless.

The loss fraction, f_{loss}

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, equals 0 for a completely closed lower boundary. If f_{loss} is set to 1, all outflow from the lower zone is treated as a loss. Water that flows out of the lower zone through D_{loss} is quite literally 'lost' forever. Physically, the loss term could represent water that is either lost to deep groundwater systems (that do not necessarily follow catchment boundaries), or even groundwater extraction wells. When using the model, it is suggested to use f_{loss} with some care; start with a value of zero, and only use any other value if it is impossible to get satisfactory results by adjusting the other calibration

parameters. At each time step, the amounts of water in the upper and lower zone are updated for the in- and outgoing fluxes, i.e.:

$$UZ_t = UZ_{t-1} + D_{2,gw} - D_{uz,lz} - Q_{uz}$$

$$LZ_t = LZ_{t-1} + D_{uz,lz} - Q_{lz} - D_{loss}$$

Note that these equations are again valid for the permeable fraction of the pixel only: storage in the direct runoff fraction equals 0 for both UZ and LZ.

Water available for infiltration and direct runoff

In the permeable fraction of each pixel $(1 - f_{dr})$, the **amount of water** that is **available for infiltration**, W_{av} $[mm]$ equals (Supit *et al.*,1994):

$$W_{av} = R \cdot \Delta t + M + D_{int} - Int$$

where:

R: Rainfall $[\frac{mm}{day}]$ M: Snow melt $[mm]$ D_{int}: Leaf drainage $[mm]$ Int: Interception $[mm]$ Δt: time step $[days]$

Since no infiltration can take place in each pixel's 'direct runoff fraction', **direct runoff** is calculated as:

$$R_d = f_{dr} \cdot W_{av}$$

where R_d is in mm per time step. Note here that W_{av} is valid for the permeable fraction only, whereas R_d is valid for the direct runoff fraction.

Routing (horizontal flow processes)

Routing of surface runoff to channel

Surface runoff is routed to the nearest downstream channel using a 4-point implicit finite-difference solution of the kinematic wave equations (Chow, 1988). The basic equations used are the equations of continuity and momentum. The continuity equation is:

$$\frac{\partial Q_{sr}}{\partial x} + \frac{\partial A_{sr}}{\partial t} = q_{sr}$$

where Q_{sr} is the surface runoff $[\frac{m^3}{s}]$, A_{sr} is the cross-sectional area of the flow $[m^2]$ and q_{sr} is the amount of lateral inflow per unit flow length $[\frac{m^2}{s}]$. The momentum equation is defined as:

$$\rho \cdot g \cdot A_{sr} \cdot (S_0 - S_f) = 0$$

where ρ is the density of the flow $[\frac{kg}{m^3}]$, g is the gravity acceleration $[\frac{m}{s^2}]$, S_0 is the topographical gradient and S_f is the friction gradient. From the momentum equation it follows that $S_0 = S_f$, which means that for the kinematic wave equations it is assumed that the water surface is parallel to the topographical surface. The continuity equation can also be written in the following finite-difference form (please note that for the sake of readability the 'sr' subscripts are omitted here from Q, A and q):

$$\frac{Q_{i+1}^{j+1} - Q_i^{j+1}}{\Delta x} + \frac{A_{i+1}^{j+1} - A_{i+1}^{j}}{\Delta t} = \frac{q_{i+1}^{j+1} - q_{i+1}^{j}}{2}$$

where j is a time index and i a space index (such that *i=1* for the most upstream cell, *i=2* for its downstream neighbor, etcetera). The momentum equation can also be expressed as (Chow et al., 1988):

$$A_{sr} = \alpha_{k,sr} \cdot Q_{sr}^{\beta_k}$$

Substituting the right-hand side of this expression in the finite-difference form of the continuity equation gives a nonlinear implicit finite-difference solution of the kinematic wave:

$$\frac{\Delta t}{\Delta x} \cdot Q_{i+1}^{j+1} \alpha_k \cdot (Q_{i+1}^{j+1})^{\beta_k} = \frac{\Delta t}{\Delta x} \cdot Q_i^{j+1} \alpha_k \cdot (Q_{i+1}^{j})^{\beta_k} \Delta t \cdot (\frac{q_{i+1}^{j+1} + q_{i+1}^{j}}{2})$$

If $_{k,sr}$ and $_k$ are known, this non-linear equation can be solved for each pixel and during each time step using an iterative procedure. This numerical solution scheme is available as a built-in function in the PCRaster software. The coefficients $_{k,sr}$ and $_k$ are calculated by substituting Manning's equation in the right-hand side of Equation:

$$A_{sr} = (\frac{n \cdot P_{sr}^{2/3}}{\sqrt{S_0}}) \cdot Q_{sr}^{3/5}$$

where n is Manning's roughness coefficient and P_{sr} is the wetted perimeter of a cross-section of the surface flow. Substituting the right-hand side of this equation for A_{sr} in equation gives:

$$\alpha_{k,sr} = (\frac{n \cdot P_{sr}^{2/3}}{\sqrt{S_0}})^{0.6}; \beta_k = 0.6$$

At present, LISFLOOD uses values for $_{k,sr}$ which are based on a static (reference) flow depth, and a flow width that equals the pixel size, Δx. For each time step, all runoff that is generated (R_s) is added as side-flow (q_{sr}). For each flowpath, the routing stops at the first downstream pixel that is part of the channel network. In other words, the routine only routes the surface runoff *to* the nearest channel; no runoff *through* the channel network is simulated at this stage (runoff- and channel routing are completely separated).

Routing of sub-surface runoff to channel

All water that flows out of the upper- and lower groundwater zone is routed to the nearest downstream channel pixel within one time step. Recalling once more that the groundwater equations are valid for the pixel's permeable fraction only, the contribution of each pixel to the nearest channel is made up of $(f_{forest} + f_{other)} \cdot (Q_u z + Q_l z)$. Figure 2.12 illustrates the routing procedure: for each pixel that contains a river channel, its contributing pixels are defined by the drainage network. For every 'river pixel' the groundwater outflow that is generated by its upstream pixels is simply summed. For instance, there are two flow paths that are contributing to the second 'river pixel' from the left in Figure below. Hence, the amount of water that is transported to this pixel equals the sum of the amounts of water produced by these flowpaths, q1 + q2. Note that, as with the surface runoff routing, no water is routed *through* the river network at this stage.

Figure: *Routing of groundwater to channel network. Groundwater flow is routed to the nearest 'channel' pixel.*

Channel routing

Flow through the channel is simulated using the kinematic wave equations. The basic equations and the numerical solution are identical to those used for the surface runoff routing:

$$\frac{\partial Q_{ch}}{\partial x} \cdot \frac{\partial A_{ch}}{\partial t} = q_{ch}$$

where Q_{ch} is the channel discharge $[\frac{m^3}{s}]$, A_{ch} is the cross-sectional area of the flow $[m^2]$ and q_ch is the amount of lateral inflow per unit flow length $[\frac{m^2}{s}]$. The momentum equation then becomes:

$$\rho \cdot g A_{ch} \cdot (S_0 - S_f) = 0$$

where S_0 now equals the gradient of the channel bed, and $S_0 = S_f$. As with the surface runoff, values for parameter $_{k,ch}$ are estimated using Manning's equation:

$$\alpha_{k,ch} = (\frac{n \cdot P_{ch}^{2/3}}{\sqrt{S_0}})^{0.6}; \beta_k = 0.6$$

At present, LISFLOOD uses values for $_{k,ch}$ which are based on a static (reference) channel flow depth (half bankfull) and measured channel dimensions. The term q_{ch} (sideflow) now represents the runoff that enters the channel per unit channel length:

$$q_{ch} = \frac{\sum Q_{sr} + \sum Q_{uz} + \sum Q_{lz} + Q_{in} + Q_{res}}{L_{ch}}$$

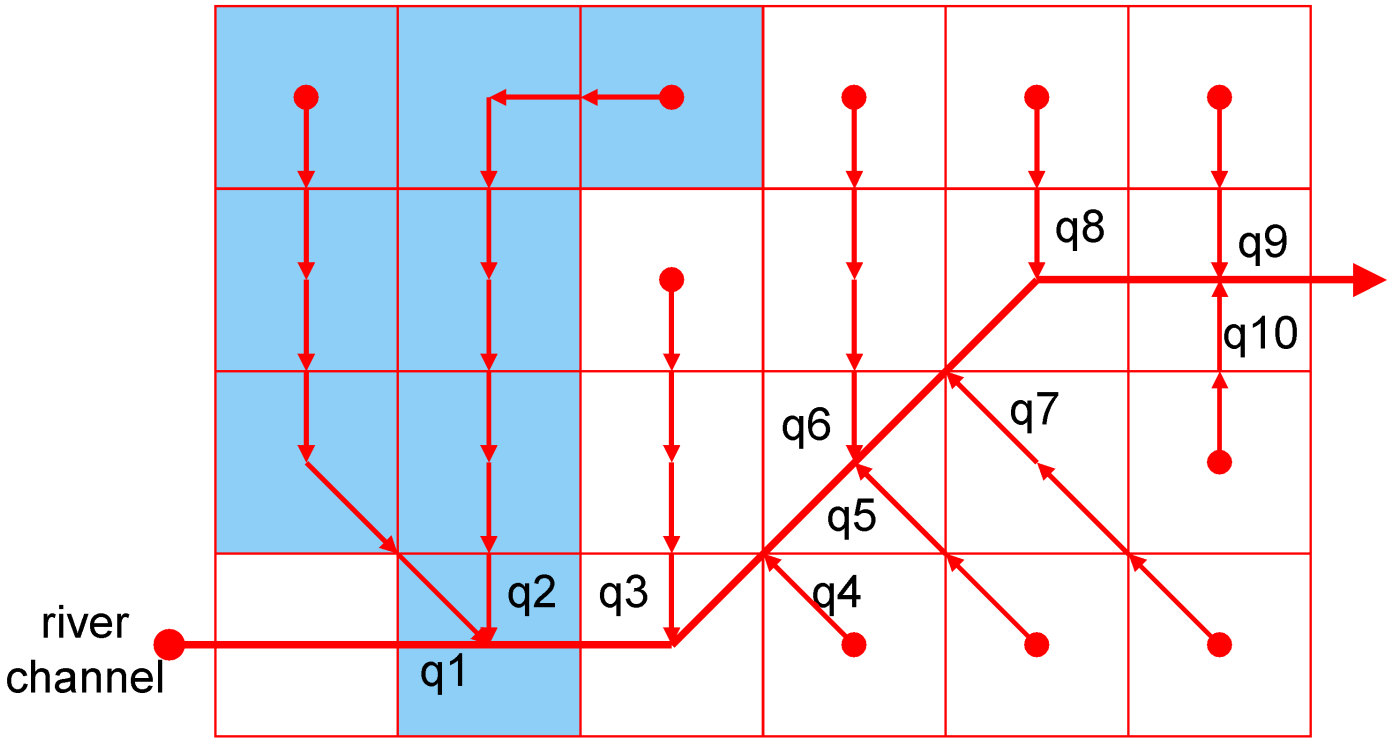

Figure 5:

Here, Q_{sr}, Q_{uz} and Q_{lz} denote the contributions of surface runoff, outflow from the upper zone and outflow from the lower zone, respectively. Q_{in} is the inflow from an external inflow hydrograph; by default its value is 0, unless the 'inflow hydrograph' option is activated (see Annex 2). Q_{res} is the water that flows out of a reservoir into the channel; by default its value is 0, unless the 'reservoir' option is activated (see **Section XXXXX**). $Q_{sr}, Q_{uz}, Q_{lz}, Q_{in}$ and Q_{res} are all expressed in [m^3 per time step. L_{ch} is the channel length $[m]$, which may exceed the pixel size (Δx) in case of meandering channels. The kinematic wave channel routing can be run using a smaller time-step than the over simulation timestep, Δt, if needed.

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Infiltration capacity

The infiltration capacity of the soil is estimated using the widely-used Xinanjiang (also known as VIC/ARNO) model (e.g. Zhao & Lui, 1995; Todini, 1996). This approach assumes that the fraction of a grid cell that is contributing to surface runoff (read: saturated) is related to the total amount of soil moisture, and that this relationship can be described through a non-linear distribution function. For any grid cell, if w_1 is the total moisture storage in the upper soil layer and w_{s1} is the maximum storage, the corresponding **saturated fraction** A_s is approximated by the following distribution function:

$$A_s = 1 - (1 - \frac{w_1}{w_{s1}})^b$$

where w_{s1} and w_1 are the maximum and actual amounts of moisture in the upper soil layer, respectively $[mm]$, and b is an empirical shape parameter. In the LISFLOOD implementation of the Xinanjiang model, A_s is defined as a fraction of the permeable fraction of each pixel (i.e. as a fraction of $(1 - d_{rf})$). The **potential infiltration capacity** $INF_{pot}[mm]$ is a function of w_s and A_s:

$$INF_{pot} = \frac{w_{s1}}{b+1} - \frac{w_{s1}}{b+1} \cdot [1 - (1 - A_s)^{\frac{b+1}{b}}]$$

Note that the shape parameter b is related to the heterogeneity within each grid cell. For a totally homogeneous grid cell b approaches zero, which reduces the above equations to a simple 'overflowing bucket' model. Before any water is draining from the soil to the groundwater zone the soil has to be completely filled up. See also red line in the Figure below: e.g. a

soil of 60% soil moisture has 40% potential infiltration capacity. A b value of 1.0 (see black line) is comparable to a leaking bucket : e.g. a soil of 60% soil moisture has only 10% potential infiltration capacity while 30% is draining directly to groundwater.

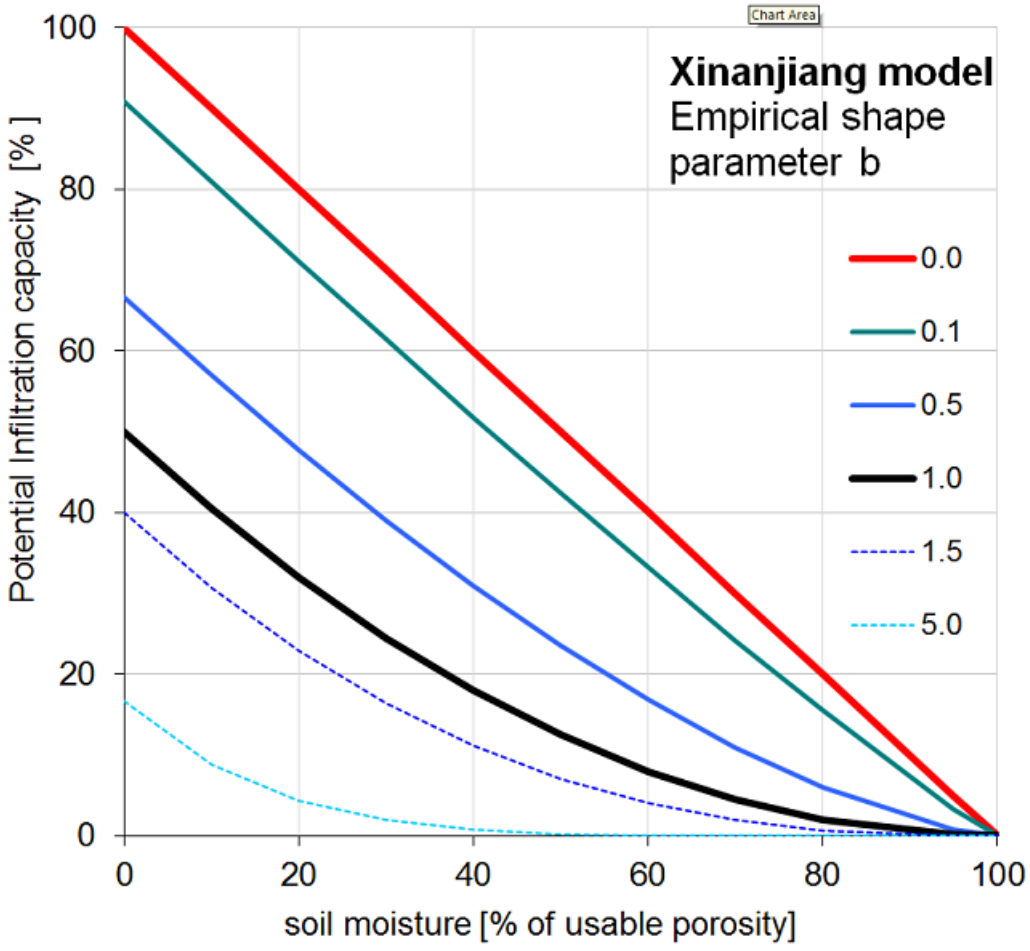

Figure 6:

Figure: *Soil moisture and potential infiltration capacity relation.*

Increasing b even further than 1 is comparable to a sieve (see figure below). Most of the water is going directly to groundwater and the potential infiltration capacity is going toward 0.

Figure: *Analogy picture of increasing Xinanjiang empirical shape parameter b.*

Interception

Interception is estimated using the following storage-based equation (Aston, 1978, Merriam, 1960):

$$Int = S_{max} \cdot [1 - e^{\frac{-k \cdot R \cdot \Delta t}{S_{max}}}]$$

where $Int[mm]$ is the interception per time step, $S_{max}[mm]$ is the maximum interception, R is the rainfall intensity $[\frac{mm}{day}]$ and the factor k accounts for the density of the vegetation. S_{max} is calculated using an empirical equation (Von Hoyningen-Huene, 1981):

$$\begin{cases} S_{max} = 0.935 + 0.498 \cdot LAI - 0.00575 \cdot LAI^2 & [LAI > 0.1] \\ S_{max} = 0 & [LAI \leq 0.1] \end{cases}$$

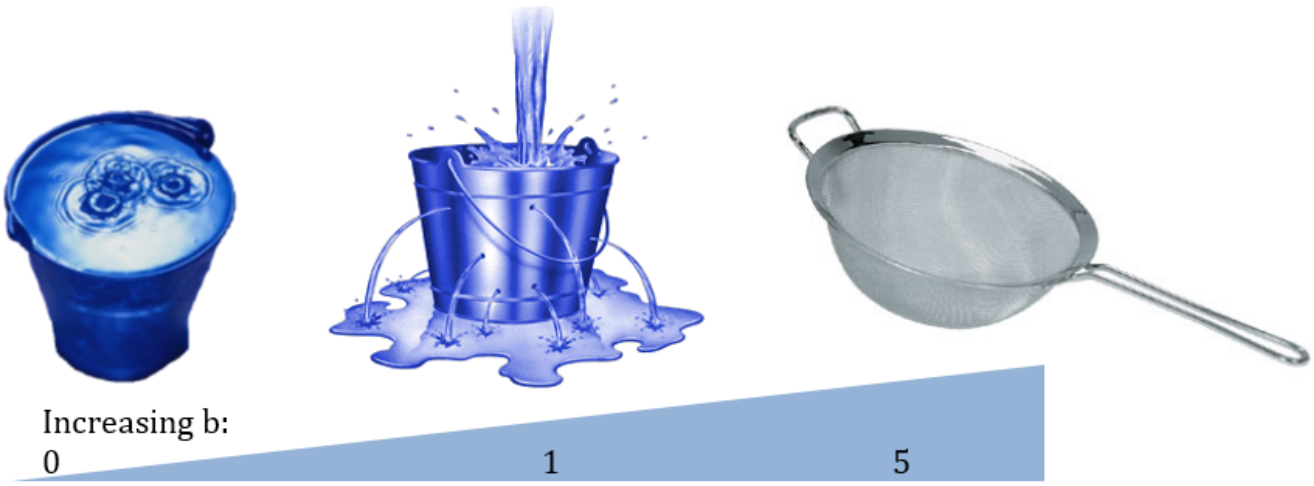

Figure 7:

where LAI is the average Leaf Area Index $[\frac{m^2}{m^2}]$ of each model element (pixel). k is estimated as:

$$k = 0.046 \cdot LAI$$

The value of Int can never exceed the interception storage capacity, which is defined as the difference between S_{max} and the accumulated amount of water that is stored as interception, Int_{cum}.

Overview

LISFLOOD model scheme

The figure below provides a first overview on the processes included in LISFLOOD:

Figure: Overview of the LISFLOOD model. P: precipitation; E: evaporation & evapotranspiration; $SnCoef$: snow melt; b_{xin}: infiltration; $Chan_{N2}$: surface runoff; GW_{perc}: drainage from upper- to lower groundwater zone; T_{uz}: outflow from upper groundwater zone; T_{lz}: outflow from lower groundwater zone; R_{ch}: drainage from subsoil to upper groundwater zone; drainage from top- to subsoil; C_{pref}: preferential flow to upper groundwater zone.

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The standard LISFLOOD model setup is made up of the following components:

- a 3-layer soil water balance sub-model
- sub-models for the simulation of groundwater and subsurface flow (using 2 parallel interconnected linear reservoirs)
- a sub-model for the routing of surface runoff to the nearest river channel
- a sub-model for the routing of channel flow

The processes that are simulated by the model include also snow melt, infiltration, interception of rainfall, leaf drainage, evaporation and water uptake by vegetation, surface runoff, preferential flow (bypass of soil layer), exchange of soil moisture between the two soil layers and drainage to the groundwater, sub-surface and groundwater flow, and flow through river channels.

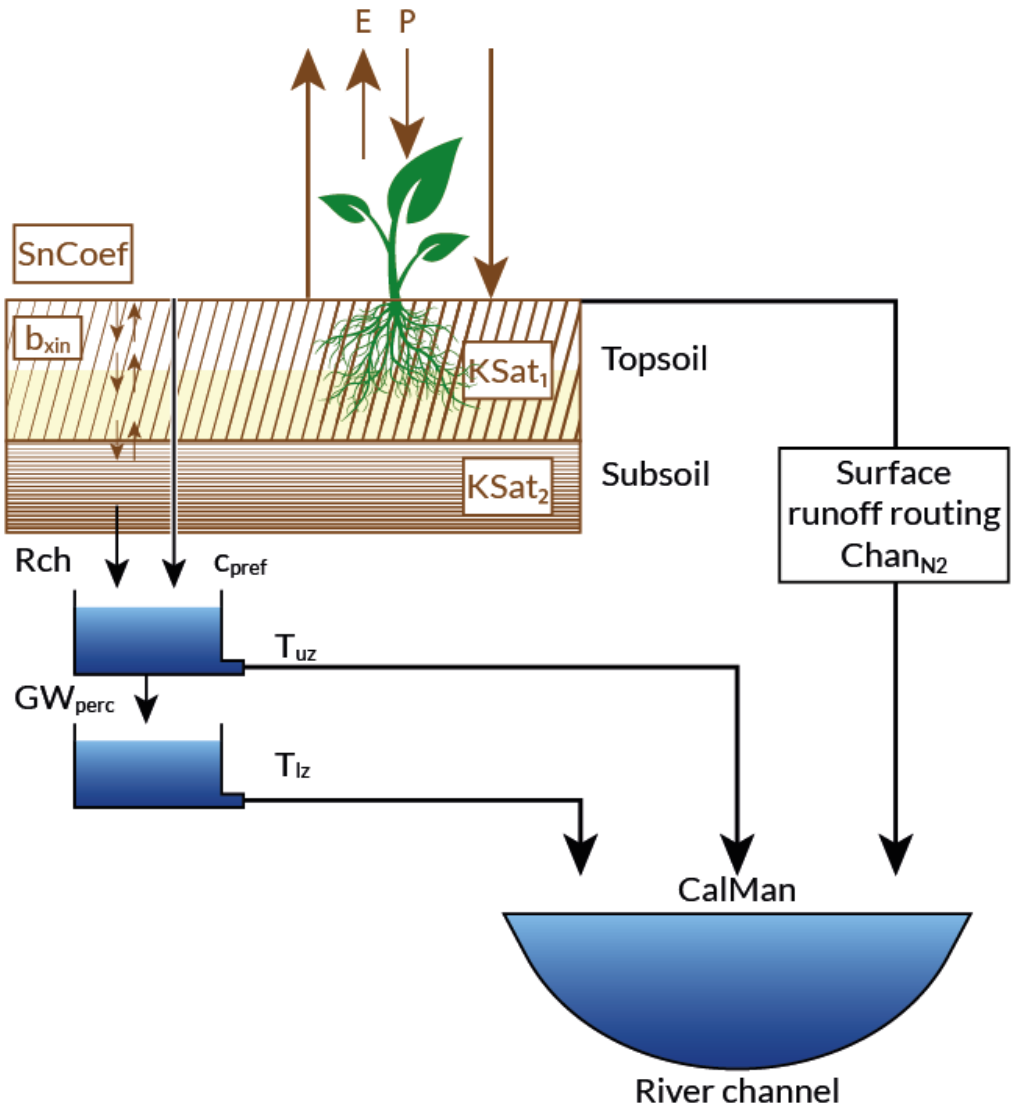
E
P
SnCoef
b_{xin}
$KSat_1$
Topsoil
$KSat_2$
Subsoil
Surface runoff routing $Chan_{N2}$
Rch
c_{pref}
T_{uz}
GW_{perc}
T_{lz}
CalMan
River channel

Figure 8:

Sub-grid variability

Before going into detail with the individual hydrological processes, here first some explaination on a larger conceptual approach on how LISFLOOD is dealing with sub-grid variability in land cover and the consecutive influence on various processes.

Representation of land cover

In LISFLOOD a number of parameters are linked directly to land cover classes. In the past, this was done through lookup tables. The spatially dominant land use class had been used (see Figure below) to assign the corresponding grid parameter values. This implies that some of the sub-grid variability in land use, and consequently in the parameter of interest, were lost.

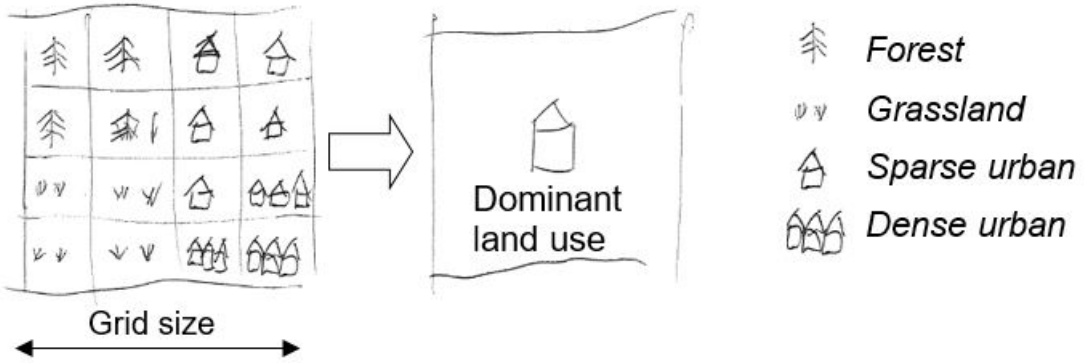

Figure 9:

Figure: *Land cover aggregation approach in previous versions of LISFLOOD.*

In order to account properly for land use dynamics, some conceptual changes have been made to render LISFLOOD more land-use sensitive. To account for the sub-grid variability in land use, we model the within-grid variability. In the latest version of the hydrological model, the spatial distribution and frequency of each class is defined as a percentage of the whole represented area of the new pixel. Combining land cover classes and modeling aggregated classes, is known as the concept of hydrological response units (HRU). The logic behind this approach is that the non-linear nature of the rainfall-runoff processes on different land cover surfaces observed in reality will be better captured. This concept is also used in models such as SWAT (Arnold and Fohrer, 2005) and PREVAH (Viviroli et al., 2009). LISFLOOD has been transferred a HRU approach on sub-grid level, as shown here:

Figure: *LISFLOOD land cover aggregation by modelling aggregated land use classes separately: Percentages of forest (dark green); water (blue), impervious surface (red), other classes (light green).*

Soil model

If a part of a pixel is made up of built-up areas this will influence that pixel's water-balance. LISFLOOD's 'direct runoff fraction's parameter (f_{dr}) defines the fraction of a pixel that is impervious.

For impervious areas, LISFLOOD assumes that: 1. A depression storage is filled by precipitation and snowmelt and emptied by evaporation 2. Any water that is not filling the depression storage, reaches the surface is added directly to surface runoff 3. The storage capacity of the soil is zero (i.e. no soil moisture storage in the direct runoff fraction) 4. There is no groundwater storage

For open water (e.g. lakes, rivers) the water fraction parameter (f_{water}) defines the fraction that is covered with water (large lakes that are in direct connection with major river channels can be modelled using LISFLOOD's lake option.

For water covered areas, LISFLOOD assumes that: 1. The loss of actual evaporation is equal to the potential evaporation on open water 2. Any water that is not evaporated, reaches the surface is added directly to surface runoff 3. The storage capacity of the soil is zero (i.e. no soil moisture storage in the water fraction) 4. There is no groundwater storage

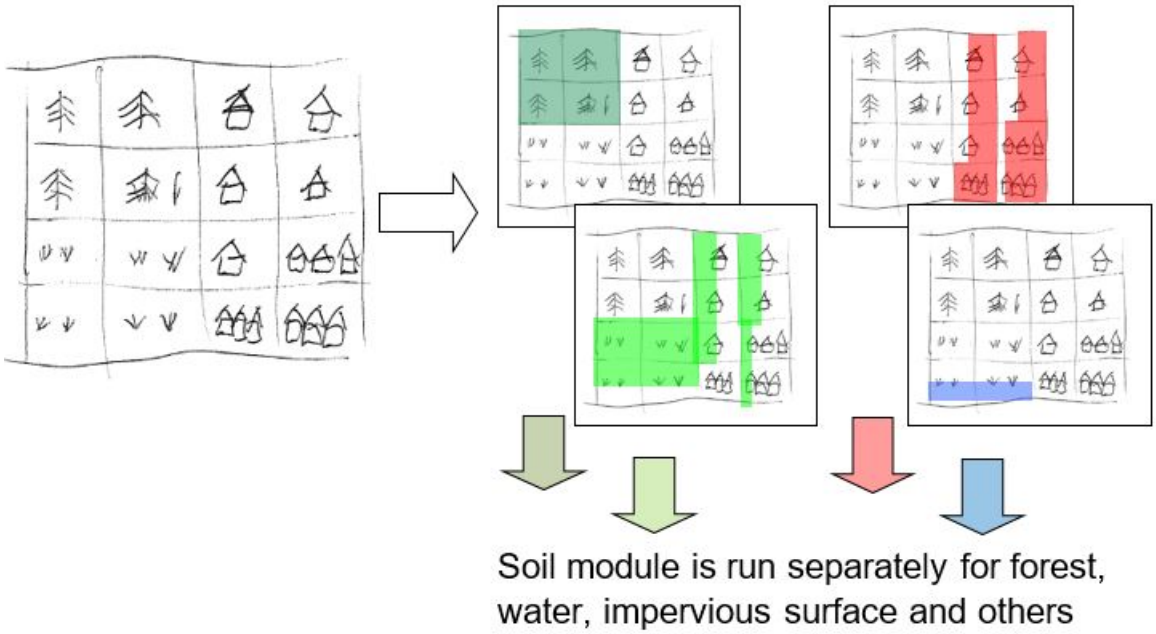

Figure 10:

For forest (f_{forest}) **or other land cover** ($f_{other} = 1 - f_{forest} - f_{dr} - f_{water}$) the description of all soil- and groundwater-related processes below (evaporation, transpiration, infiltration, preferential flow, soil moisture redistribution and groundwater flow) are valid. While the modelling structure for forest and other classes is the same, the difference is the use of different map sets for leaf area index, soil and soil hydraulic properties. Because of the nonlinear nature of the rainfall runoff processes this should yield better results than running the model with average parameter values. The table below summarises the profiles of the four individually modelled categories of land cover classes.

Table: *Summary of hydrological properties of the four categories modelled individually in the modified version of LISFLOOD.*

Category	Evapotranspiration	Soil	Runoff
Forest	High level of evapo-transpiration (high Leaf area index) seasonally dependent	Large rooting depth	Low concentration time
Impervious surface	Not applicable	Not applicable	Surface runoff but initial loss and depression storage, fast concentration time
Inland water	Maximum evaporation	Not applicable	Fast concentration time
Other (agricultural areas, non-forested natural area, pervious surface of urban areas)	Evapotranspiration lower than for forest but still significant	Rooting depth lower than for forest but still significant	Medium concentration time

If you activate any of LISFLOOD's options for writing internal model fluxes to time series or maps (described in **Chapter XXX - AD I think this chapter is still missing!**), the model will report the real fluxes, which are the fluxes multiplied by the corresponding fraction. The Figure below illustrates this for evapotranspiration (evaporation and transpiration) which calculated differently for each of this four aggregated classes. The total sum of evapotranspiration for a pixel is calculated by adding up the fluxes for each class multiplied by the fraction of each class.

Figure: $ET_{forest} \rightarrow ET_{other} \rightarrow ET_{dr} \rightarrow ET_{water}$ *simulation of aggregated land cover classes in LISFLOOD.*

In this example, evapotranspiration (ET) is simulated for each aggregated class separately ($ET_{forest}, ET_{dr}, ET_{water}, ET_{other}$) As result of the soil model you get four different surface fluxes weighted by the corresponding fraction ($f_{dr}, f_{water}, f_{forest}, f_{other}$), respectively two fluxes for the upper and lower groundwater zone and for groundwater loss also weighted by the corresponding fraction (f_{forest}, f_{other}). However a lot of the internal flux or states (e.g. preferential flow for forested areas)

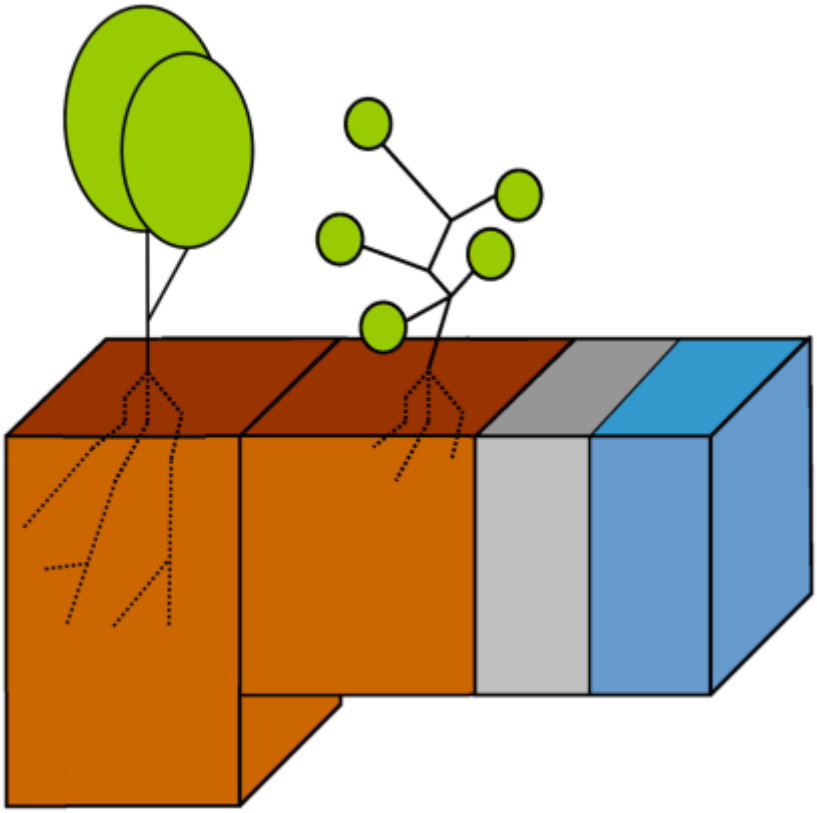

Figure 11:

Note that the energy that has been 'consumed' already for the evaporation of intercepted water is simply subtracted here in order to respect the overall energy balance.

The **actual transpiration rate** is reduced when the amount of moisture in the soil is small. In the model, a reduction factor is applied to simulate this effect:

$$r_{WS} = \frac{w_1 - w_{wp1}}{w_{crit1} - w_{wp1}}$$

where w_1 is the amount of moisture in the upper soil layer $[mm]$, $w_{wp1}[mm]$ is the amount of soil moisture at wilting point (pF 4.2) and $w_{crit1}[mm]$ is the amount of moisture below which water uptake is reduced and plants start closing their stomata. The **critical amount of soil moisture** is calculated as:

$$w_{crit1} = (1 - p) \cdot (w_{fc1} - w_{wp1}) + w_{wp1}$$

where $w_{fc1}[mm]$ is the amount of soil moisture at field capacity and p is the soil water depletion fraction. R_{WS} varies between 0 and 1. Negative values and values greater than 1 are truncated to 0 and 1, respectively. p represents the fraction of soil moisture between w_{fc1} and w_{wp1} that can be extracted from the soil without reducing the transpiration rate. Its value is a function of both vegetation type and the potential evapotranspiration rate. The procedure to estimate p is described in detail in Supit & Van Der Goot (2003). The following Figure illustrates the relation between $R_{WS}, w, w_{crit}, w_{fc}, w_{wp}$:

Figure: *Reduction of transpiration in case of water stress. r_{ws} decreases linearly to zero between w_{crit} and w_{wp}.*

The **actual transpiration** T_a is now calculated as:

$$T_a = r_{WS} \cdot T_{max}$$

with T_a and T_{max} in $[mm]$.

Transpiration is set to zero when the soil is frozen (i.e. when frost index F exceeds its critical threshold). The amount of **moisture in the upper soil layer** is updated after the transpiration calculations:

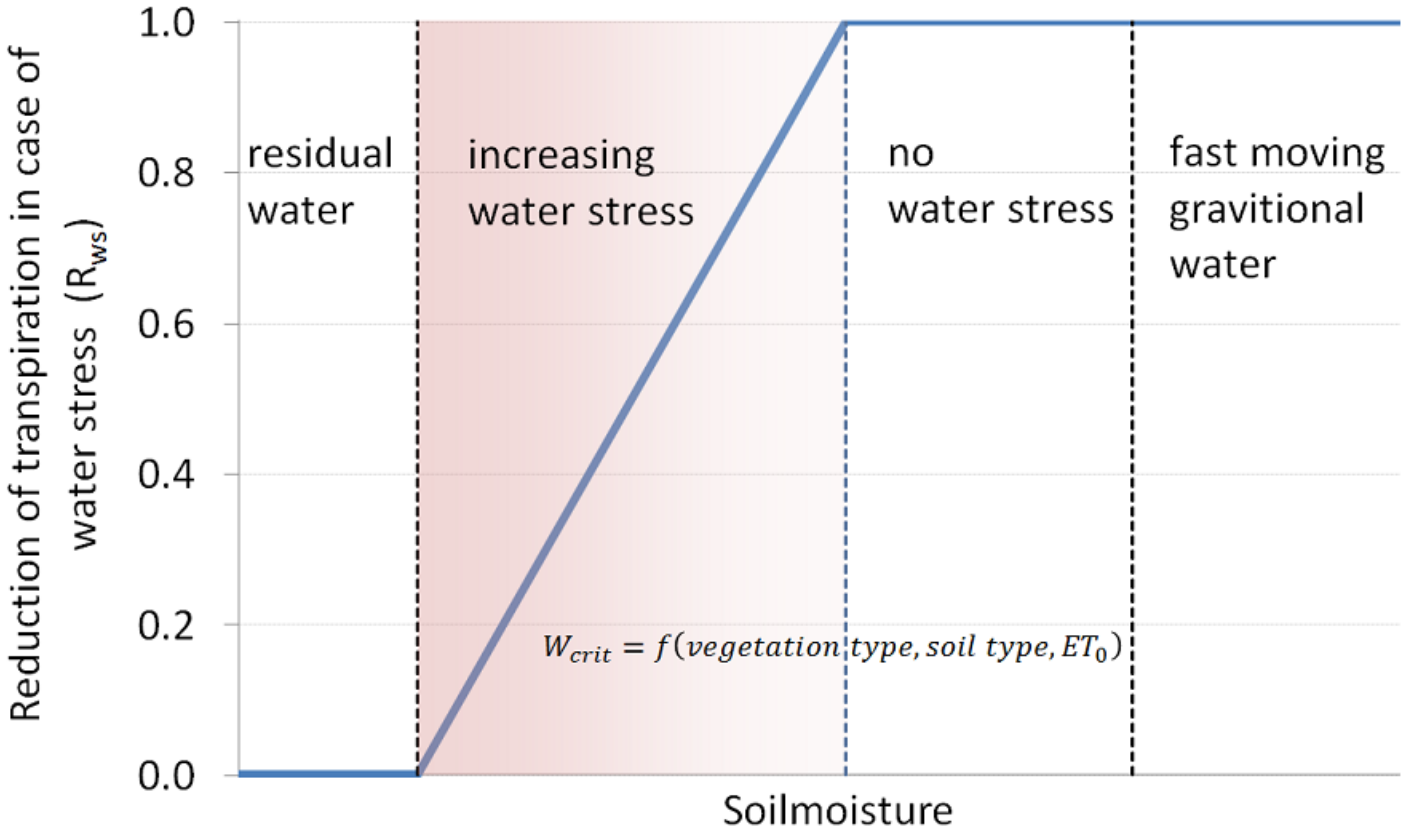

Figure 12:

$$w_1 = w_1 - T_a$$

Preferential bypass flow

For the simulation of preferential bypass flow –i.e. flow that bypasses the soil matrix and drains directly to the groundwater- no generally accepted equations exist. Because ignoring preferential flow completely will lead to unrealistic model behavior during extreme rainfall conditions, a very simple approach is used in LISFLOOD. During each time step, a fraction of the water that is available for infiltration is added to the groundwater directly (i.e. without first entering the soil matrix). It is assumed that this fraction is a power function of the relative saturation of the topsoil, which results in an equation that is somewhat similar to the excess soil water equation used in the HBV model (e.g. Lindström *et al.*, 1997):

$$D_{pref,gw} = W_{av} \cdot (\frac{w_1}{w_{s1}})^{c_{pref}}$$

where $D_{pref,gw}$ is the amount of preferential flow per time step $[mm]$, W_{av} is the amount of water that is available for infiltration, and c_{pref} is an empirical shape parameter. f_{dr} is the 'direct runoff fraction'

–

, which is the fraction of each pixel that is made up by urban area and open water bodies (i.e. preferential flow is only simulated in the permeable fraction of each pixel) . The equation results in a preferential flow component that becomes increasingly important as the soil gets wetter.

The Figure below shows with $c_{pref} = 0$ (red line) every available water for infiltration is converted into preferential flow and bypassing the soil. $c_{pref} = 1$ (black line) gives a linear relation e.g. at 60% soil saturation 60% of the available water is bypassing the soil matrix. With increasing c_{pref} the percentage of preferential flow is decreasing.

Figure: *Soil moisture and preferential flow relation.*

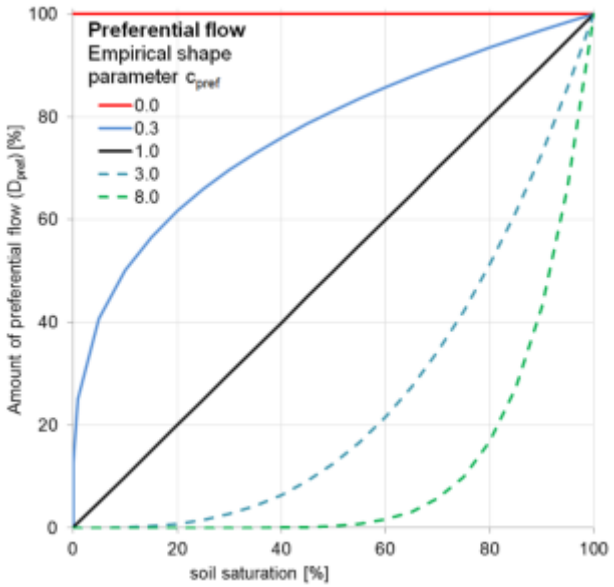

Figure 13:

Rain and snow

If the average temperature is below 1°C, all precipitation is assumed to be snow. A snow correction factor is used to correct for undercatch of snow precipitation. Unlike rain, snow accumulates on the soil surface until it melts. The rate of snowmelt is estimated using a simple degree-day factor method. Degree-day factor type snow melt models usually take the following form (e.g. see WMO, 1986):

$$M = C_m(T_{avg} - T_m)$$

where M is the rate of snowmelt, T_{avg} is the average daily temperature, T_m is some critical temperature and C_m is a degree-day factor $[\frac{mm}{°C\ day}]$.

Speers *et al.* (1979) developed an extension of this equation which accounts for accelerated snowmelt that takes place when it is raining (cited in Young, 1985). The equation is supposed to apply when rainfall is greater than 30 mm in 24 hours. Moreover, although the equation is reported to work sufficiently well in forested areas, it is not valid in areas that are above the tree line, where radiation is the main energy source for snowmelt). LISFLOOD uses a variation on the equation of Speers *et al.* The modified equation simply assumes that for each mm of rainfall, the rate of snowmelt increases with 1% (compared to a 'dry' situation). This yields the following equation:

$$M = C_m \cdot C_{Seasonal}(1 + 0.01 \cdot R\Delta t)(T_{avg} - T_m) \cdot \Delta t$$

where M is the snowmelt per time step $[mm]$, R is rainfall (not snow!) intensity $[\frac{mm}{day}]$, and Δt is the time interval $[days]$. T_m has a value of 0 $°C$, and C_m is a degree-day factor $[\frac{mm}{°C \cdot day}]$.

However, it should be stressed that the value of C_m can actually vary greatly both in space and time (e.g. see Martinec *et al.*, 1998). Therefore, **in practice this parameter is often treated as a calibration constant**. A low value of C_m indicates slow snow melt. $C_{Seasonal}$ is a seasonal variable melt factor which is also used in several other models (e.g. Anderson 2006, Viviroli et al., 2009). There are mainly two reasons to use a seasonally variable melt factor:

- The solar radiation has an effect on the energy balance and varies with the time of the year.
- The albedo of the snow has a seasonal variation, because fresh snow is more common in the mid winter and aged snow in the late winter/spring. This produce an even greater seasonal variation in the amount of net solar radiation

The following Figure shows an example where a mean value of: 3.0 $\frac{mm}{°C \cdot day}$ is used. The value of C_m is reduced by 0.5 at 21^{st} December and a 0.5 is added on the 21^{st} June. In between a sinus function is applied

Figure: *Sinus shaped snow melt coefficient (C_m) as a function of days of year.*

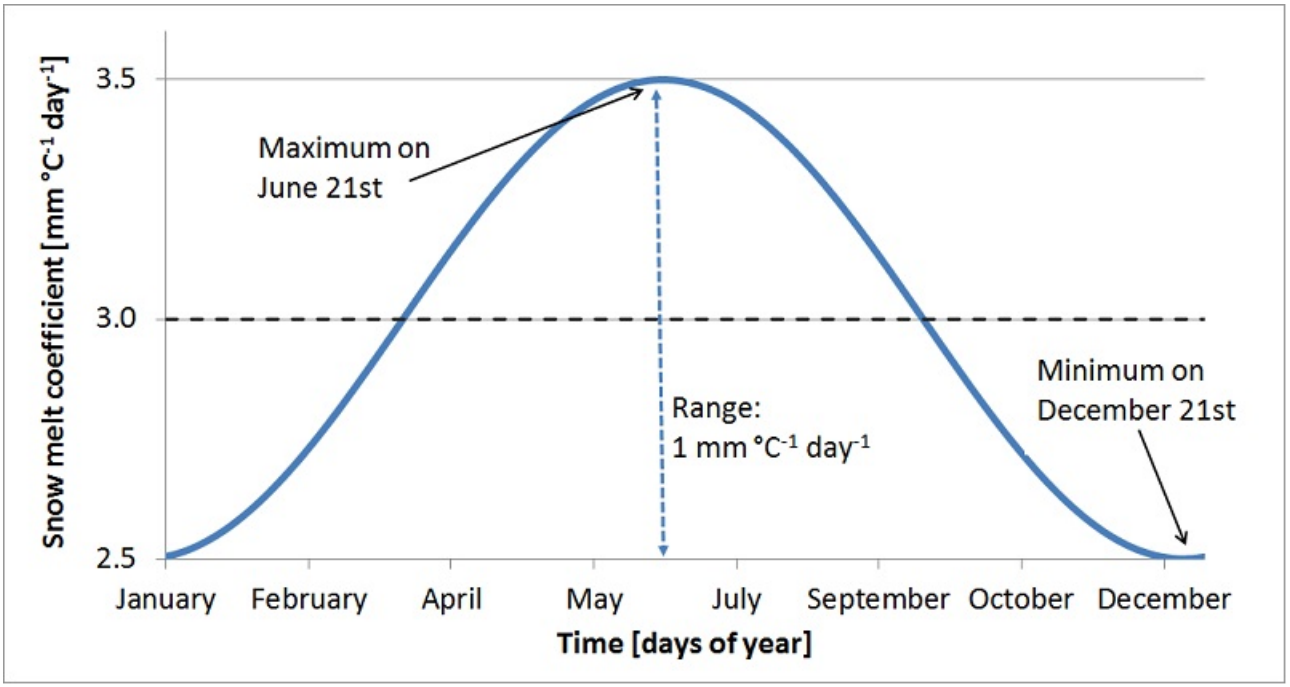

Figure 14:

At high altitudes, where the temperature never exceeds $1°C$, the model accumulates snow without any reduction because of melting loss. In these altitudes runoff from glacier melt is an important part. The snow will accumulate and converted into firn. Then firn is converted to ice and transported to the lower regions. This can take decades or even hundred years. In the ablation area the ice is melted. In LISFLOOD this process is emulated by melting the snow in higher altitudes on an annual basis over summer. A sinus function is used to start ice melting in summer (from 15 June till 15 September) using the temperature of zone B:

Figure: *Sinus shaped ice melt coefficient as a function of days of year.*

The amount of snowmelt and ice melt together can never exceed the actual snow cover that is present on the surface.

For large pixel sizes, there may be considerable sub-pixel heterogeneity in snow accumulation and melt, which is a particular problem if there are large elevation differences within a pixel. Because of this, snow melt and accumulation are modelled separately for 3 separate elevation zones, which are defined at the sub-pixel level. This is shown in Figure below:

Figure: *Definition of sub-pixel elevation zones for snow accumulation and melt modelling. Snowmelt and accumulation calculations in each zone are based on elevation (and derived temperature) in centroid of each zone.*

The division in elevation zones was changed from a uniform distribution in the previous LISFLOOD version to a normal distribution, which fits better to the real distribution of e.g. 100m SRTM DEM pixels in a 5x5km grid cell. Three elevation zones A, B, and C are defined with each zone occupying one third of the pixel surface. Assuming further that T_{avg} is valid for the average pixel elevation, average temperature is extrapolated to the centroids of the lower (A) and upper (C) elevation zones, using a fixed temperature lapse rate, L, of 0.0065 °C per meter elevation difference. Snow, snowmelt and snow accumulation are subsequently modelled separately for each elevation zone, assuming that temperature can be approximated by the temperature at the centroid of each respective zone.

Direct evaporation from the soil surface

The **maximum amount of evaporation from the soil surface** equals the maximum evaporation from a shaded soil surface, $ES_{max}[mm]$, which is computed as:

$$ES_{max} = ES0 \cdot e^{\left(\frac{-\kappa_{gb} \cdot LAI}{\Delta t}\right)}$$

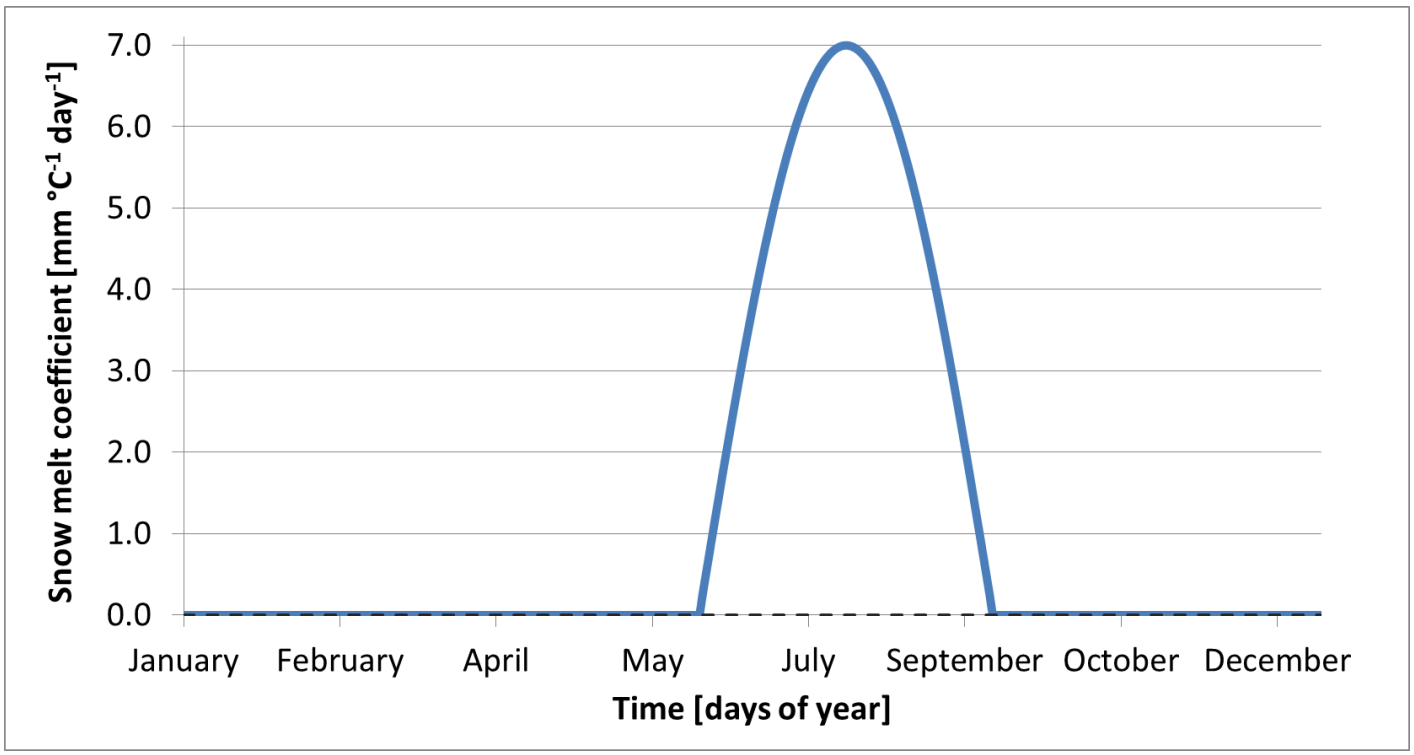

Figure 15:

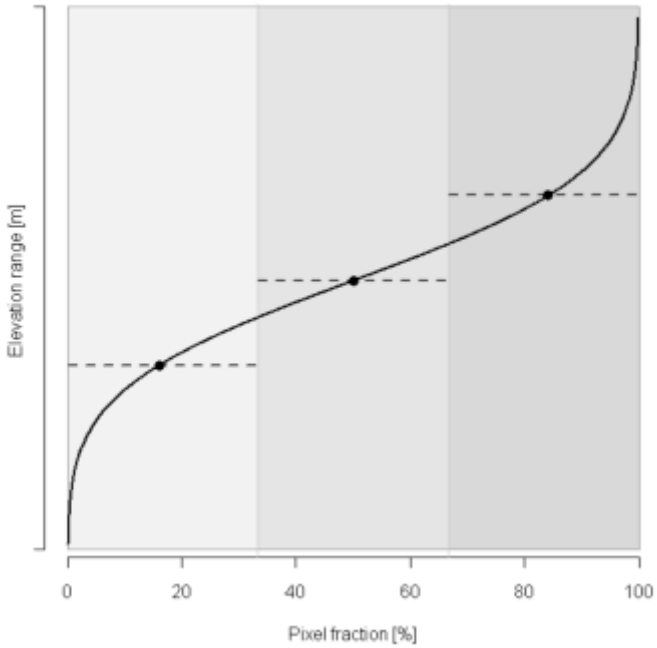

Figure 16:

where $ES0$ is the potential evaporation rate from bare soil surface $[\frac{mm}{day}]$. The **actual evaporation from the soil** mainly depends on the amount of soil moisture near the soil surface: evaporation decreases as the topsoil is drying. In the model this is simulated using a reduction factor which is a function of the number of days since the last rain storm (Stroosnijder, 1987, 1982):

$$ES_a = ES_{max} \cdot (\sqrt{D_{slr}} - \sqrt{D_{slr} - 1})$$

The variable D_{slr} represents the number of days since the last rain event. Its value accumulates over time: if the amount of water that is available for infiltration (W_{av}) remains below a critical threshold it increases by an amount of $\Delta t[days]$ for each time step. It is reset to 1 only if the critical amount of water is exceeded (In the LISFLOOD settings file this critical amount is currently expressed as an *intensity* $[\frac{mm}{day}]$. This is because the equation was originally designed for a daily time step only. Because the current implementation will likely lead to *DSLR* being reset too frequently, the exact formulation may change in future versions (e.g. by keeping track of the accumulated available water of the last 24 hours).

The **actual soil evaporation** is always the smallest value out of the result of the equation above and the available amount of moisture in the soil, i.e.:

$$ES_a = \min(ES_a, w_1 - w_{res1})$$

where $w_1[mm]$ is the amount of moisture in the upper soil layer and $w_{res1}[mm]$ is the residual amount of soil moisture. Like transpiration, direct evaporation from the soil is set to zero if the soil is frozen. The amount of moisture in the upper soil layer is updated after the evaporation calculations:

$$w_1 = w_1 - ES_a$$

Soil model

If a part of a pixel is made up of built-up areas this will influence that pixel's water-balance. LISFLOOD's 'direct runoff fraction's parameter (f_{dr}) defines the fraction of a pixel that is impervious. **For impervious areas**, LISFLOOD assumes that:

1. A depression storage is filled by precipitation and snowmelt and emptied by evaporation
2. Any water that is not filling the depression storage, reaches the surface is added directly to surface runoff
3. The storage capacity of the soil is zero (i.e. no soil moisture storage in the direct runoff fraction)
4. There is no groundwater storage

For open water (e.g. lakes, rivers) the water fraction parameter (f_{water}) defines the fraction that is covered with water (large lakes that are in direct connection with major river channels can be modelled using LISFLOOD's lake option. **For water covered areas**, LISFLOOD assumes that:

1. The loss of actual evaporation is equal to the potential evaporation on open water
2. Any water that is not evaporated, reaches the surface is added directly to surface runoff
3. The storage capacity of the soil is zero (i.e. no soil moisture storage in the water fraction)
4. There is no groundwater storage

For the part of a pixel that is **forest** (f_{forest}) **or other land cover** ($f_{other} = 1 - f_{forest} - f_{dr} - f_{water}$) the description of all soil- and groundwater-related processes below (evaporation, transpiration, infiltration, preferential flow, soil moisture redistribution and groundwater flow) are valid. While the modelling structure for forest and other classes is the same, the difference is the use of different map sets for leaf area index, soil and soil hydraulic properties. Because of the nonlinear nature of the rainfall runoff processes this should yield better results than running the model with average parameter values. The table below summarises the profiles of the four individually modelled categories of land cover classes.

Table: *Summary of hydrological properties of the four categories modelled individually in the modified version of LISFLOOD.*

Category	Evapotranspiration	Soil	Runoff
Forest	High level of evapo-transpiration (high Leaf area index) seasonally dependent	Large rooting depth	Low concentration time
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Inland water	Maximum evaporation	Not applicable	Fast concentration time
Other (agricultural areas, non-forested natural area, pervious surface of urban areas)	Evapotranspiration lower than for forest but still significant	Rooting depth lower than for forest but still significant	Medium concentration time

If you activate any of LISFLOOD's options for writing internal model fluxes to time series or maps (described in OPTIONAL LISFLOOD PROCESSES AND OUTPUT), the model will report the real fluxes, which are the fluxes multiplied by the corresponding fraction. The Figure below illustrates this for evapotranspiration (evaporation and transpiration) which calculated differently for each of this four aggregated classes. The total sum of evapotranspiration for a pixel is calculated by adding up the fluxes for each class multiplied by the fraction of each class.

Figure *$ET_{forest} \rightarrow ET_{other} \rightarrow ET_{dr} \rightarrow ET_{water}$ simulation of aggregated land cover classes in LISFLOOD.*

In this example, evapotranspiration (ET) is simulated for each aggregated class separately ($ET_{forest}, ET_{dr}, ET_{water}, ET_{other}$) As result of the soil model you get four different surface fluxes weighted by the corresponding fraction ($f_{dr}, f_{water}, f_{forest}, f_{other}$), respectively two fluxes for the upper and lower groundwater zone and for groundwater loss also weighted by the corresponding fraction (f_{forest}, f_{other}). However a lot of the internal flux or states (e.g. preferential flow for forested areas) can be written to disk as map or timeseries by activate LISFLOOD's options.

Soil moisture redistribution

The description of the moisture fluxes out of the subsoil (and also between the upper- and lower soil layer) is based on the simplifying assumption that the flow of soil moisture is entirely gravity-driven. Starting from **Darcy's law for 1-D vertical flow rate**:

$$q = -K(\theta) \cdot [\frac{\partial h(\theta)}{\partial z} - 1]$$

where $q[\frac{mm}{day}]$ is the flow rate out of the soil (e.g. upper soil layer, lower soil layer); $K(\theta)[\frac{mm}{day}]$ is the hydraulic conductivity (as a function of the volumetric moisture content of the soil, $\theta[\frac{mm^3}{mm^3}]$ and $\frac{\partial h(\theta)}{\partial z}$ is the matric potential gradient. If we assume a matric potential gradient of zero, the equation reduces to:

$$q = K(\theta)$$

This implies a flow that is always in downward direction, at a rate that equals the conductivity of the soil. The relationship between hydraulic conductivity and soil moisture status is described by the **Van Genuchten equation** (van Genuchten, 1980), here re-written in terms of mm water slice, instead of volume fractions:

$$K(w) = K_s \cdot \sqrt{(\frac{w - w_r}{w_s - w_r})} \cdot \{1 - [1 - (\frac{w - w_r}{w_s - w_r})^{\frac{1}{m}}]^m\}^2$$

where K_s is the **saturated conductivity** of the soil $[\frac{mm}{day}]$, and w, w_r and w_s are the actual, residual and maximum amounts of moisture in the soil respectively (all in $[mm]$). Parameter m is calculated from the pore-size index, λ (which is related to soil texture):

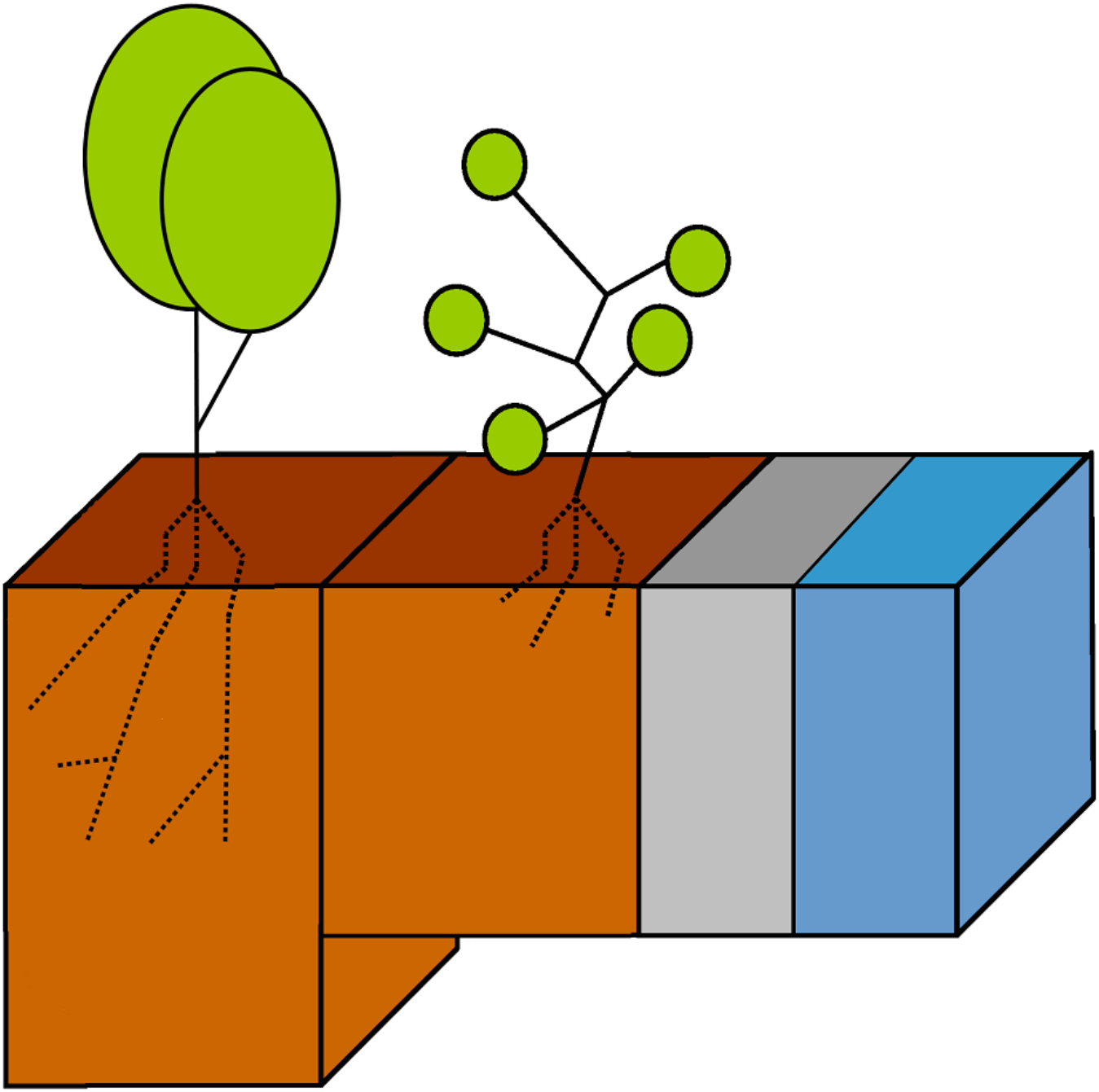

Figure 17:

$$m = \frac{\lambda}{\lambda + 1}$$

For large values of Δt (e.g. 1 day) the above equation often results in amounts of outflow that exceed the available soil moisture storage, i.e:

$$K(w)\Delta t > w - w_r$$

In order to solve the soil moisture equations correctly an iterative procedure is used. At the beginning of each time step, the conductivities for both soil layers $[K_1(w_1), K_2(w_2)]$ are calculated using the Van Genuchten equation. Multiplying these values with the time step and dividing by the available moisture gives a Courant-type numerical stability indicator for each respective layer:

$$C_1 = \frac{K_1(w_1) \cdot \Delta t}{w_1 - w_{r1}}$$

$$C_2 = \frac{K_2(w_2) \cdot \Delta t}{w_2 - w_{r2}}$$

A Courant number that is greater than 1 implies that the calculated outflow exceeds the available soil moisture, resulting in loss of mass balance. Since we need a stable solution for both soil layers, the **'overall' Courant number** for the soil moisture routine is the largest value out of C_1 and C_2:

$$C_{soil} = max(C_1, C_2)$$

In principle, rounding C_{soil} up to the nearest integer gives the number sub-steps needed for a stable solution. In practice, it is often preferable to use a critical Courant number that is lower than 1, because high values can result in unrealistic 'jumps' in the simulated soil moisture pattern when the soil is near saturation (even though mass balance is preserved). Hence, making the critical Courant number a user-defined value C_{crit}, the number of sub-steps becomes:

$$SubSteps = roundup(\frac{C_{soil}}{C_{crit}})$$

and the corresponding sub-time-step, $\Delta' t$:

$$\Delta' t = \frac{\Delta t}{SubSteps}$$

In brief, the iterative procedure now involves the following steps. First, the number of sub-steps and the corresponding sub-time-step are computed as explained above. The amounts of soil moisture in the upper and lower layer are copied to temporary variables

$$w'_1$$

and

$$w'_2$$

. Two variables,

$$D_{1,2}$$

(flow from upper to lower soil layer) and

$$D_{2,gw}$$

(flow from lower soil layer to groundwater) are initialized (set to zero). Then, for each sub-step, the following sequence of calculations is performed:

1. compute hydraulic conductivity for both layers
2. compute flux from upper to lower soil layer for this sub-step ($D'_{1,2}$, can never exceed storage capacity in lower layer):

$$D'_{1,2} = min[K_1(w'_1)\Delta t, w'_{s2} - w'_2]$$

3. compute flux from lower soil layer to groundwater for this sub-step ($D'_{2,gw}$), can never exceed available water in lower layer):

$$D'_{2,gw} = min[K_2(w'_2)\Delta t, w'_2 - w'_{r2}]$$

4. update w'_1 and w'_2
5. add

$$D'_{1,2}$$

to

$$D_{1,2}$$

; add

$$D'_{2,gw}$$

to

$$D_{2,gw}$$

If the soil is frozen ($F >$ critical threshold), both $D_{1,2}$ and $D_{2,gw}$ are set to zero. After the iteration loop, the amounts of soil moisture in both layers are updated as follows:

$$w_1 = w_1 - D_{1,2}$$

$$w_2 = w_2 + D_{1,2} - D_{2,gw}$$

Sub-grid variability

Before going into detail with the individual hydrological processes, here first some explaination on a larger conceptual approach on how LISFLOOD is dealing with sub-grid variability in land cover and the consecutive influence on various processes.

Representation of land cover

In LISFLOOD a number of parameters are linked directly to land cover classes. In the past, this was done through lookup tables. The spatially dominant land use class had been used (see Figure below) to assign the corresponding grid parameter values. This implies that some of the sub-grid variability in land use, and consequently in the parameter of interest, were lost.

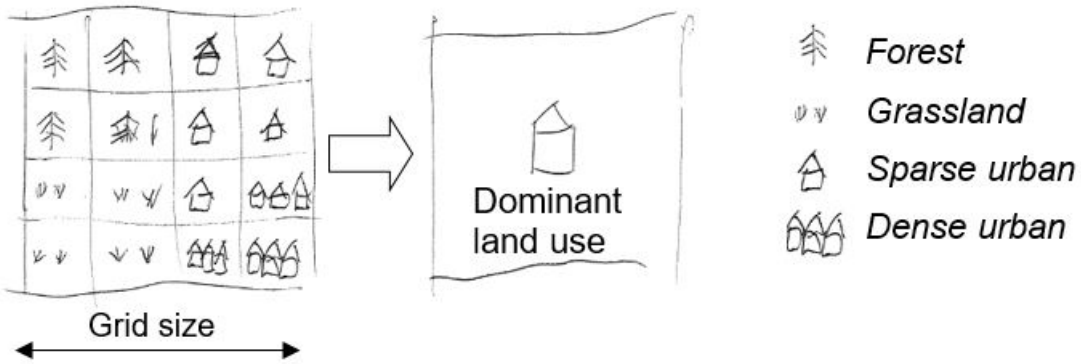

Figure 18:

Figure: *Land cover aggregation approach in previous versions of LISFLOOD.*

In order to account properly for land use dynamics, some conceptual changes have been made to render LISFLOOD more land-use sensitive. To account for the sub-grid variability in land use, we model the within-grid variability. In the latest version of the hydrological model, the spatial distribution and frequency of each class is defined as a percentage of the whole represented area of the new pixel. Combining land cover classes and modeling aggregated classes, is known as the concept of hydrological response units (HRU). The logic behind this approach is that the non-linear nature of the rainfall-runoff processes on different land cover surfaces observed in reality will be better captured. This concept is also used in models such as SWAT (Arnold and Fohrer, 2005) and PREVAH (Viviroli et al., 2009). LISFLOOD has been transferred a HRU approach on sub-grid level, as shown here:

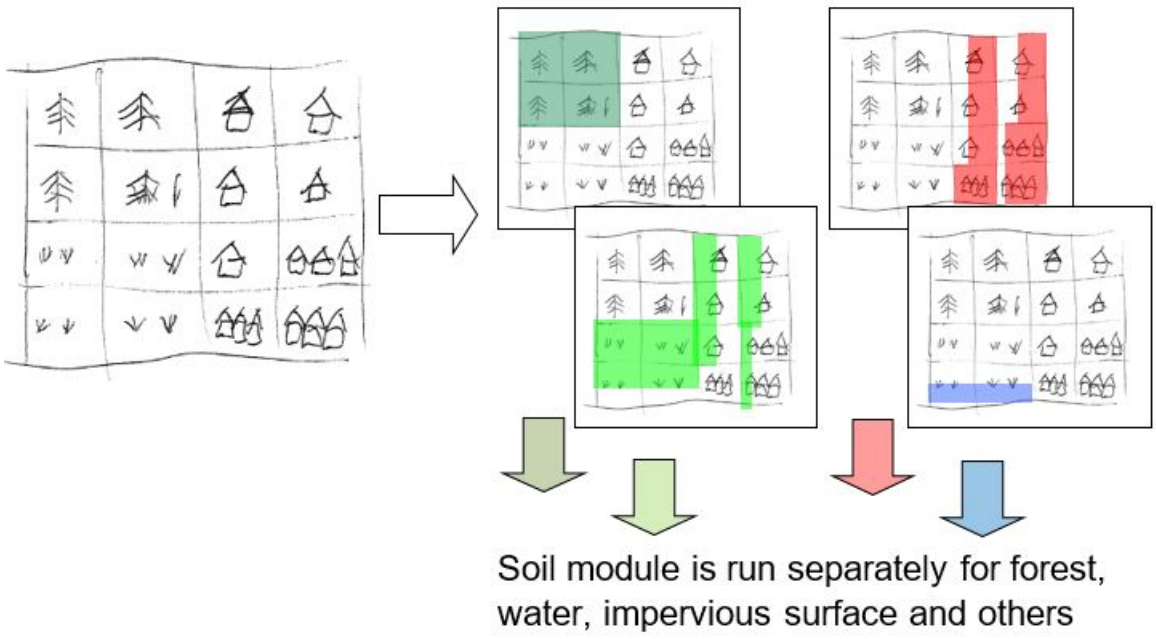

Figure 19:

Figure: *LISFLOOD land cover aggregation by modelling aggregated land use classes separately: Percentages of forest (dark green); water (blue), impervious surface (red), other classes (light green).*

Soil model

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For impervious areas, LISFLOOD assumes that: 1. A depression storage is filled by precipitation and snowmelt and emptied by evaporation 2. Any water that is not filling the depression storage, reaches the surface is added directly to surface runoff 3. The storage capacity of the soil is zero (i.e. no soil moisture storage in the direct runoff fraction) 4. There is no groundwater storage

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runoff processes this should yield better results than running the model with average parameter values. The table below summarises the profiles of the four individually modelled categories of land cover classes.

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If you activate any of LISFLOOD's options for writing internal model fluxes to time series or maps (described in **Chapter XXX - AD I think this chapter is still missing!**), the model will report the real fluxes, which are the fluxes multiplied by the corresponding fraction. The Figure below illustrates this for evapotranspiration (evaporation and transpiration) which calculated differently for each of this four aggregated classes. The total sum of evapotranspiration for a pixel is calculated by adding up the fluxes for each class multiplied by the fraction of each class.

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Routing of sub-surface runoff to channel

All water that flows out of the upper- and lower groundwater zone is routed to the nearest downstream channel pixel within one time step. Recalling once more that the groundwater equations are valid for the pixel's permeable fraction only, the contribution of each pixel to the nearest channel is made up of $(f_{forest} + f_{other)} \cdot (Q_u z + Q_l z)$. The Figure below illustrates the routing procedure: for each pixel that contains a river channel, its contributing pixels are defined by the drainage network. For every 'river pixel' the groundwater outflow that is generated by its upstream pixels is simply summed. For instance, there are two flow paths that are contributing to the second 'river pixel' from the left in Figure below. Hence, the amount of water that is transported to this pixel equals the sum of the amounts of water produced by these flowpaths, q1 + q2. Note that, as with the surface runoff routing, no water is routed *through* the river network at this stage.

Figure: *Routing of groundwater to channel network. Groundwater flow is routed to the nearest 'channel' pixel.*

Routing of surface runoff to channel

Surface runoff is routed to the nearest downstream channel using a 4-point implicit finite-difference solution of the kinematic wave equations (Chow, 1988). The basic equations used are the equations of continuity and momentum. The continuity equation is:

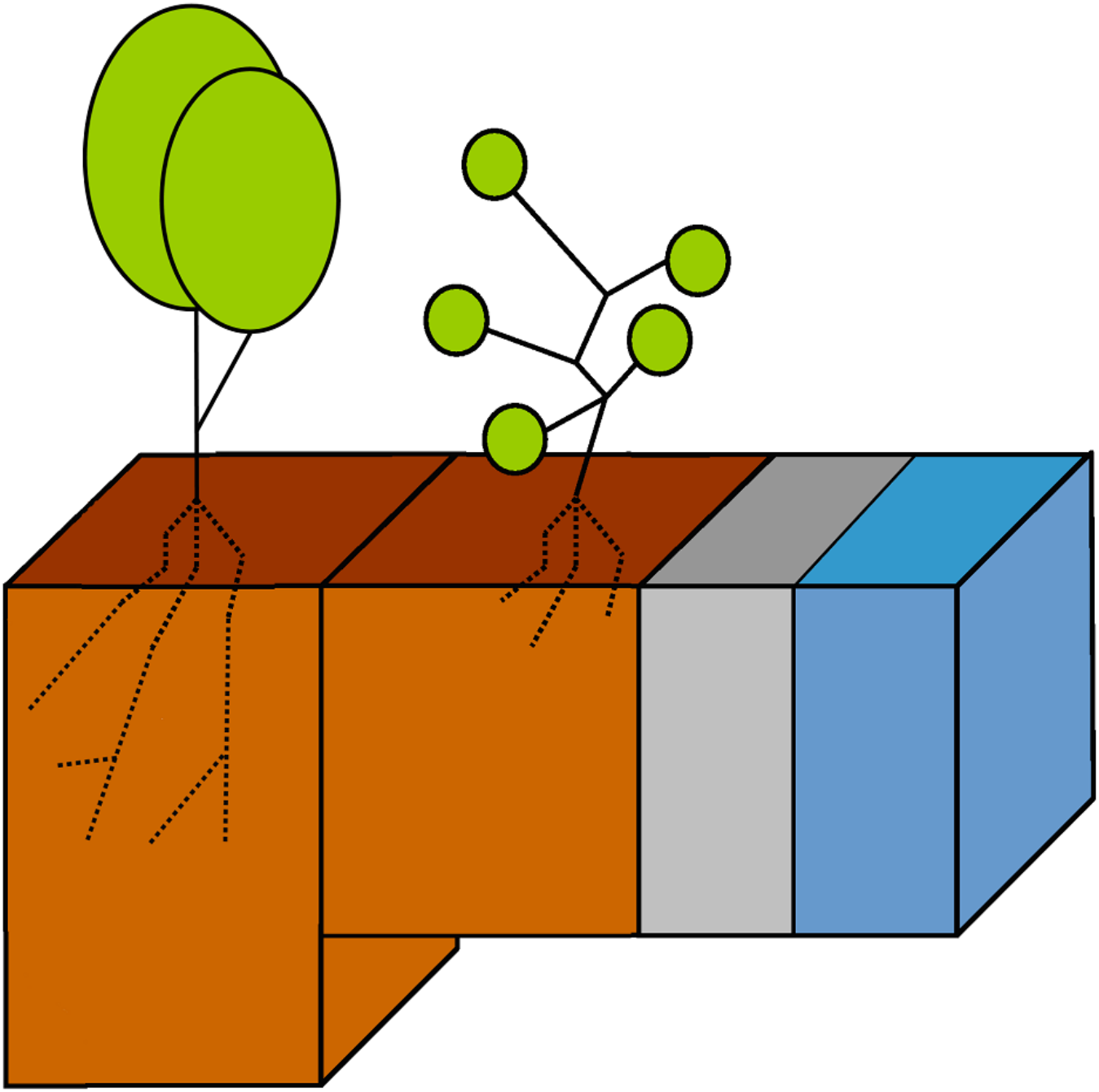


Figure 20:

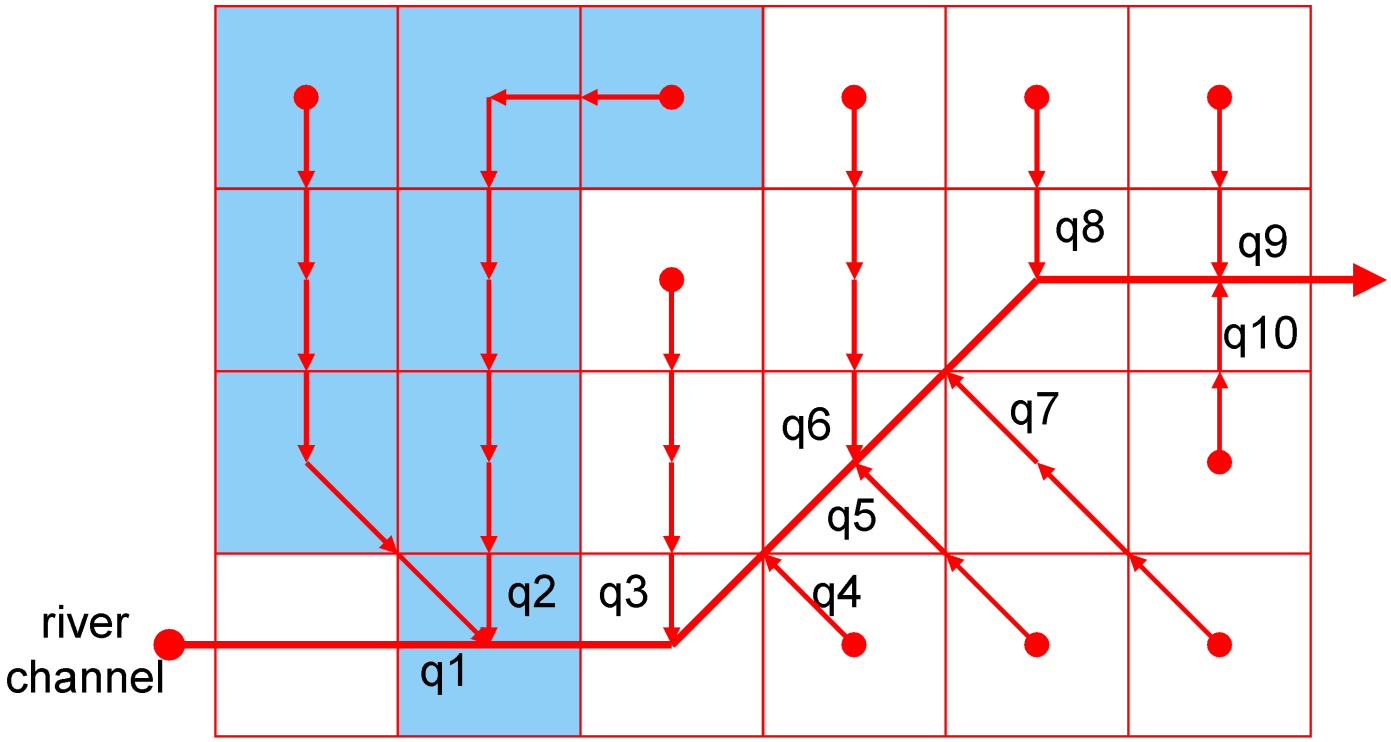


Figure 21:

$$\frac{\partial Q_{sr}}{\partial x} + \frac{\partial A_{sr}}{\partial t} = q_{sr}$$

where Q_{sr} is the surface runoff $[\frac{m^3}{s}]$, A_{sr} is the cross-sectional area of the flow $[m^2]$ and q_{sr} is the amount of lateral inflow per unit flow length $[\frac{m^2}{s}]$. The momentum equation is defined as:

$$\rho \cdot g \cdot A_{sr} \cdot (S_0 - S_f) = 0$$

where ρ is the density of the flow $[\frac{kg}{m^3}]$, g is the gravity acceleration $[\frac{m}{s^2}]$, S_0 is the topographical gradient and S_f is the friction gradient. From the momentum equation it follows that $S_0 = S_f$, which means that for the kinematic wave equations it is assumed that the water surface is parallel to the topographical surface. The continuity equation can also be written in the following finite-difference form (please note that for the sake of readability the 'sr' subscripts are omitted here from Q, A and q):

$$\frac{Q_{i+1}^{j+1} - Q_i^{j+1}}{\Delta x} + \frac{A_{i+1}^{j+1} - A_{i+1}^{j}}{\Delta t} = \frac{q_{i+1}^{j+1} - q_{i+1}^{j}}{2}$$

where j is a time index and i a space index (such that *i=1* for the most upstream cell, *i=2* for its downstream neighbor, etcetera). The momentum equation can also be expressed as (Chow et al., 1988):

$$A_{sr} = \alpha_{k,sr} \cdot Q_{sr}^{\beta_k}$$

Substituting the right-hand side of this expression in the finite-difference form of the continuity equation gives a nonlinear implicit finite-difference solution of the kinematic wave:

$$\frac{\Delta t}{\Delta x} \cdot Q_{i+1}^{j+1} \alpha_k \cdot (Q_{i+1}^{j+1})^{\beta_k} = \frac{\Delta t}{\Delta x} \cdot Q_i^{j+1} \alpha_k \cdot (Q_{i+1}^{j})^{\beta_k} \Delta t \cdot (\frac{q_{i+1}^{j+1} + q_{i+1}^{j}}{2})$$

If $_{k,sr}$ and $_k$ are known, this non-linear equation can be solved for each pixel and during each time step using an iterative procedure. This numerical solution scheme is available as a built-in function in the PCRaster software. The coefficients $_{k,sr}$ and $_k$ are calculated by substituting Manning's equation in the right-hand side of Equation:

$$A_{sr} = (\frac{n \cdot P_{sr}^{2/3}}{\sqrt{S_0}}) \cdot Q_{sr}^{3/5}$$

where n is Manning's roughness coefficient and P_{sr} is the wetted perimeter of a cross-section of the surface flow. Substituting the right-hand side of this equation for A_{sr} in equation gives:

$$\alpha_{k,sr} = (\frac{n \cdot P_{sr}^{2/3}}{\sqrt{S_0}})^{0.6}; \beta_k = 0.6$$

At present, LISFLOOD uses values for $_{k,sr}$ which are based on a static (reference) flow depth, and a flow width that equals the pixel size, Δx. For each time step, all runoff that is generated (R_s) is added as side-flow (q_{sr}). For each flowpath, the routing stops at the first downstream pixel that is part of the channel network. In other words, the routine only routes the surface runoff *to* the nearest channel; no runoff *through* the channel network is simulated at this stage (runoff- and channel routing are completely separated).

Drainage (vertical flow processes)

Frost index soil

When the soil surface is frozen, this affects the hydrological processes occurring near the soil surface. To estimate whether the soil surface is frozen or not, a frost index F is calculated. The equation is based on Molnau & Bissell (1983, cited in Maidment 1993), and adjusted for variable time steps. The rate at which the frost index changes is given by:

$$\frac{dF}{dt} = -(1 - A_f) \cdot F - T_{av} \cdot e^{-0.04 \cdot K \cdot d_s / w \cdot e_s}$$

$\frac{dF}{dt}$ is expressed in $[\frac{^{\circ}C}{day} \cdot \frac{1}{day}]$. A_f is a decay coefficient $[\frac{1}{day}]$, K is a a snow depth reduction coefficient $[\frac{1}{cm}]$, d_s is the (pixel-average) depth of the snow cover (expressed as mm equivalent water depth), and $w \cdot e_s$ is a parameter called snow water equivalent, which is the equivalent water depth water of a snow cover (Maidment, 1993). In LISFLOOD, A_f and K are set to 0.97 and 0.57 $[\frac{1}{cm}]$ respectively, and $w \cdot e_s$ is taken as 0.1, assuming an average snow density of 100 $\frac{kg}{m^3}$ (Maidment, 1993). The soil is considered frozen when the frost index rises above a critical threshold of 56. For each time step the value of F $[\frac{^{\circ}C}{day}]$ is updated as:

$$F(t) = F(t-1) + \frac{dF}{dt}\Delta t$$

Note: F is not allowed to become less than 0.

When the frost index rises above a threshold of 56, every soil process is frozen and transpiration, evaporation, infiltration and the outflow to the second soil layer and to upper groundwater layer is set to zero. Any rainfall is bypassing the soil and transformed into surface runoff till the frost index is equal or less than 56.

Water available for infiltration and direct runoff

In the permeable fraction of each pixel $(1 - f_{dr})$, the amount of water that is available for infiltration, W_{av} $[mm]$ equals (Supit *et al.*,1994):

$$W_{av} = R \cdot \Delta t + M + D_{int} - Int$$

where:

R: Rainfall $[\frac{mm}{day}]$ M: Snow melt $[mm]$ D_{int}: Leaf drainage $[mm]$ Int: Interception $[mm]$ Δt: time step $[days]$

Since no infiltration can take place in each pixel's 'direct runoff fraction', direct runoff is calculated as:

$$R_d = f_{dr} \cdot W_{av}$$

where R_d is in mm per time step. Note here that W_{av} is valid for the permeable fraction only, whereas R_d is valid for the direct runoff fraction.

Infiltration capacity

The infiltration capacity of the soil is estimated using the widely-used Xinanjiang (also known as VIC/ARNO) model (e.g. Zhao & Lui, 1995; Todini, 1996). This approach assumes that the fraction of a grid cell that is contributing to surface runoff (read: saturated) is related to the total amount of soil moisture, and that this relationship can be described through a non-linear distribution function. For any grid cell, if w_1 is the total moisture storage in the upper soil layer and w_s1 is the maximum storage, the corresponding saturated fraction A_s is approximated by the following distribution function:

$$A_s = 1 - (1 - \frac{w_1}{w_{s1}})^b$$

where w_{s1} and w_1 are the maximum and actual amounts of moisture in the upper soil layer, respectively $[mm]$, and b is an empirical shape parameter. In the LISFLOOD implementation of the Xinanjiang model, A_s is defined as a fraction of the permeable fraction of each pixel (i.e. as a fraction of $(1 - d_{rf})$). The infiltration capacity $INF_{pot}[mm]$ is a function of w_s and A_s:

$$INF_{pot} = \frac{w_{s1}}{b+1} - \frac{w_{s1}}{b+1} \cdot [1 - (1 - A_s)^{\frac{b+1}{b}}]$$

Note that the shape parameter b is related to the heterogeneity within each grid cell. For a totally homogeneous grid cell b approaches zero, which reduces the above equations to a simple 'overflowing bucket' model. Before any water is draining from the soil to the groundwater zone the soil has to be completely filled up. See also red line in the Figure below: e.g. a soil of 60% soil moisture has 40% potential infiltration capacity. A b value of 1.0 (see black line) is comparable to a leaking bucket : e.g. a soil of 60% soil moisture has only 10% potential infiltration capacity while 30% is draining directly to groundwater.

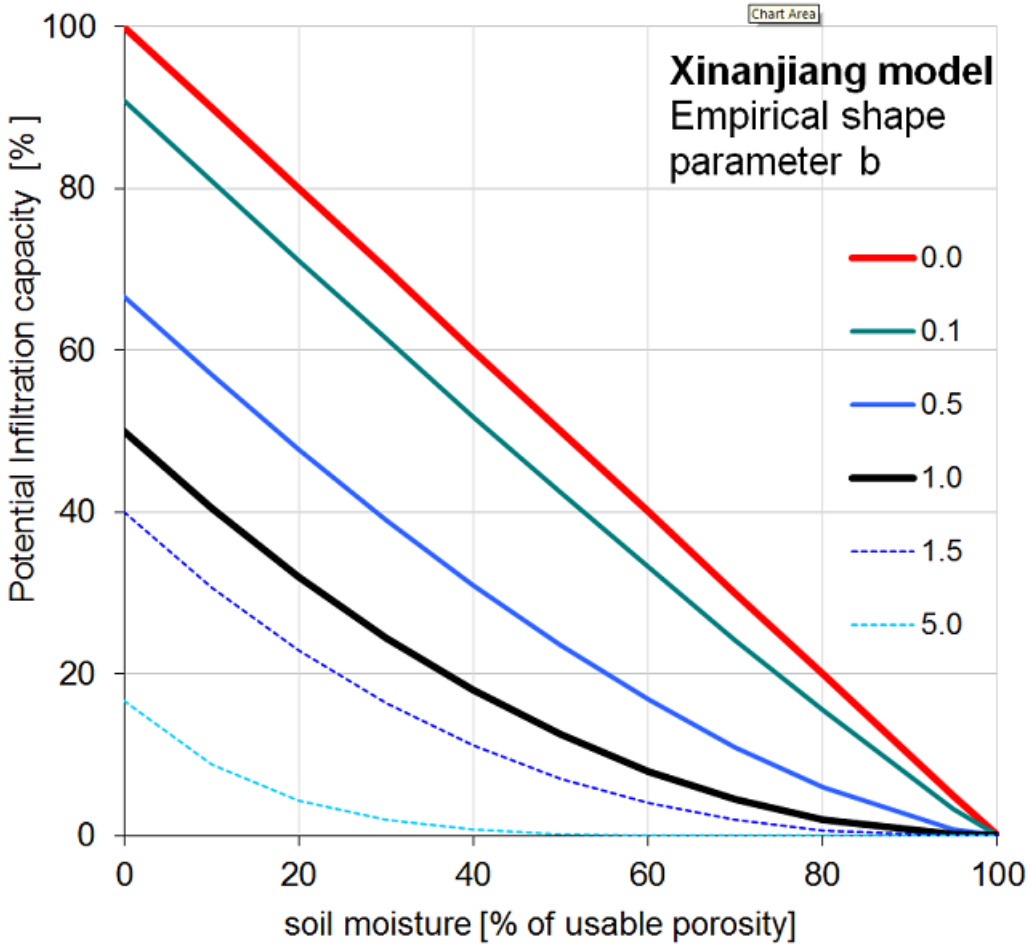


Figure 22:

Figure: *Soil moisture and potential infiltration capacity relation.*

Increasing b even further than 1 is comparable to a sieve (see figure below). Most of the water is going directly to groundwater and the potential infiltration capacity is going toward 0.

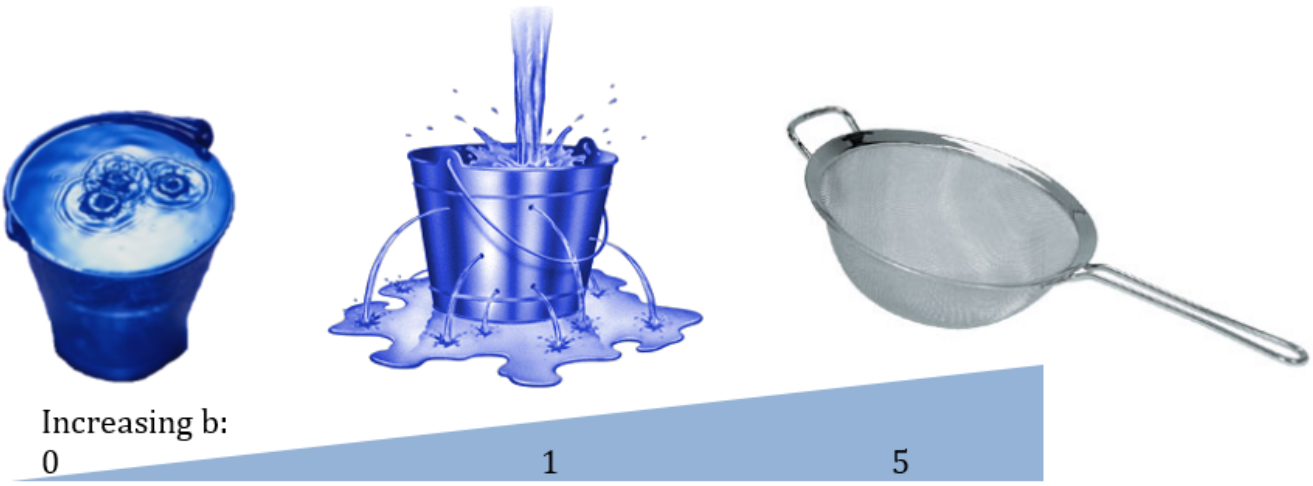


Figure 23:

Figure: *Analogy picture of increasing Xinanjiang empirical shape parameter b.*

Actual infiltration and surface runoff

The actual infiltration INF_{act} $[mm]$ is now calculated as:

$$INF_{act} = min(INF_{pot}, W_{av} - D_{pref,gw})$$

Finally, the surface runoff R_s $[mm]$ is calculated as:

$$R_s = R_d + (1 - f_{dr}) \cdot (W_{av} - D_{pref,gw} - INF_{act})$$

where R_d is the direct runoff (generated in the pixel's 'direct runoff fraction'). If the soil is frozen ($F >$ critical threshold) no infiltration takes place. The amount of moisture in the upper soil layer is updated after the infiltration calculations:

$$w_1 = w_1 + INF_{act}$$

Soil moisture redistribution

The description of the moisture fluxes out of the subsoil (and also between the upper- and lower soil layer) is based on the simplifying assumption that the flow of soil moisture is entirely gravity-driven. Starting from Darcy's law for 1-D vertical flow:

$$q = -K(\theta) \cdot [\frac{\partial h(\theta)}{\partial z} - 1]$$

where $q[\frac{mm}{day}]$ is the flow rate out of the soil (e.g. upper soil layer, lower soil layer); $K(\theta)[\frac{mm}{day}]$ is the hydraulic conductivity (as a function of the volumetric moisture content of the soil, $\theta[\frac{mm^3}{mm^3}]$ and $\frac{\partial h(\theta)}{\partial z}$ is the matric potential gradient. If we assume a matric potential gradient of zero, the equation reduces to:

$$q = K(\theta)$$

This implies a flow that is always in downward direction, at a rate that equals the conductivity of the soil. The relationship between hydraulic conductivity and soil moisture status is described by the Van Genuchten equation (van Genuchten, 1980), here re-written in terms of mm water slice, instead of volume fractions:

$$K(w) = K_s \cdot \sqrt{(\frac{w - w_r}{w_s - w_r})} \cdot \{1 - [1 - (\frac{w - w_r}{w_s - w_r})^{\frac{1}{m}}]^m\}^2$$

where K_s is the saturated conductivity of the soil $[\frac{mm}{day}]$, and w, w_r and w_s are the actual, residual and maximum amounts of moisture in the soil respectively (all in $[mm]$). Parameter m is calculated from the pore-size index, λ (which is related to soil texture):

$$m = \frac{\lambda}{\lambda + 1}$$

For large values of Δt (e.g. 1 day) the above equation often results in amounts of outflow that exceed the available soil moisture storage, i.e:

$$K(w)\Delta t > w - w_r$$

In order to solve the soil moisture equations correctly an iterative procedure is used. At the beginning of each time step, the conductivities for both soil layers $[K_1(w_1), K_2(w_2)]$ are calculated using the Van Genuchten equation. Multiplying these values with the time step and dividing by the available moisture gives a Courant-type numerical stability indicator for each respective layer:

$$C_1 = \frac{K_1(w_1) \cdot \Delta t}{w_1 - w_{r1}}$$

$$C_2 = \frac{K_2(w_2) \cdot \Delta t}{w_2 - w_{r2}}$$

A Courant number that is greater than 1 implies that the calculated outflow exceeds the available soil moisture, resulting in loss of mass balance. Since we need a stable solution for both soil layers, the 'overall' Courant number for the soil moisture routine is the largest value out of C_1 and C_2:

$$C_{soil} = max(C_1, C_2)$$

In principle, rounding C_{soil} up to the nearest integer gives the number sub-steps needed for a stable solution. In practice, it is often preferable to use a critical Courant number that is lower than 1, because high values can result in unrealistic 'jumps' in the simulated soil moisture pattern when the soil is near saturation (even though mass balance is preserved). Hence, making the critical Courant number a user-defined value C_{crit}, the number of sub-steps becomes:

$$SubSteps = roundup(\frac{C_{soil}}{C_{crit}})$$

and the corresponding sub-time-step, $\Delta' t$:

$$\Delta' t = \frac{\Delta t}{SubSteps}$$

In brief, the iterative procedure now involves the following steps. First, the number of sub-steps and the corresponding sub-time-step are computed as explained above. The amounts of soil moisture in the upper and lower layer are copied to temporary variables w'_1 and w'_2. Two variables, $D_{1,2}$ (flow from upper to lower soil layer) and $D_{2,gw}$ (flow from lower soil layer to groundwater) are initialized (set to zero). Then, for each sub-step, the following sequence of calculations is performed:

1. compute hydraulic conductivity for both layers
2. compute flux from upper to lower soil layer for this sub-step ($D'_{1,2}$, can never exceed storage capacity in lower layer):

$$D'_{1,2} = min[K_1(w'_1)\Delta t, w'_{s2} - w'_2]$$

3. compute flux from lower soil layer to groundwater for this sub-step ($D'_{2,gw}$), can never exceed available water in lower layer):

$$D'_{2,gw} = min[K_2(w'_2)\Delta t, w'_2 - w'_{r2}]$$

4. update w'_1 and w'_2
5. add $D'_{1,2}$ to $D_{1,2}$; add $D'_{2,gw}$ to $D_{2,gw}$

If the soil is frozen ($F >$ critical threshold), both $D_{1,2}$ and $D_{2,gw}$ are set to zero. After the iteration loop, the amounts of soil moisture in both layers are updated as follows:

$$w_1 = w_1 - D_{1,2}$$

$$w_2 = w_2 + D_{1,2} - D_{2,gw}$$

Preferential bypass flow

For the simulation of preferential bypass flow –i.e. flow that bypasses the soil matrix and drains directly to the groundwater- no generally accepted equations exist. Because ignoring preferential flow completely will lead to unrealistic model behavior during extreme rainfall conditions, a very simple approach is used in LISFLOOD. During each time step, a fraction of the water that is available for infiltration is added to the groundwater directly (i.e. without first entering the soil matrix). It is assumed that this fraction is a power function of the relative saturation of the topsoil, which results in an equation that is somewhat similar to the excess soil water equation used in the HBV model (e.g. Lindström *et al.*, 1997):

$$D_{pref,gw} = W_{av} \cdot (\frac{w_1}{w_{s1}})^{c_{pref}}$$

where $D_{pref,gw}$ is the amount of preferential flow per time step $[mm]$, W_{av} is the amount of water that is available for infiltration, and c_{pref} is an empirical shape parameter. f_{dr} is the 'direct runoff fraction'

–

, which is the fraction of each pixel that is made up by urban area and open water bodies (i.e. preferential flow is only simulated in the permeable fraction of each pixel) . The equation results in a preferential flow component that becomes increasingly important as the soil gets wetter.

The Figure below shows with $c_{pref} = 0$ (red line) every available water for infiltration is converted into preferential flow and bypassing the soil. $c_{pref} = 1$ (black line) gives a linear relation e.g. at 60% soil saturation 60% of the available water is bypassing the soil matrix. With increasing c_{pref} the percentage of preferential flow is decreasing.

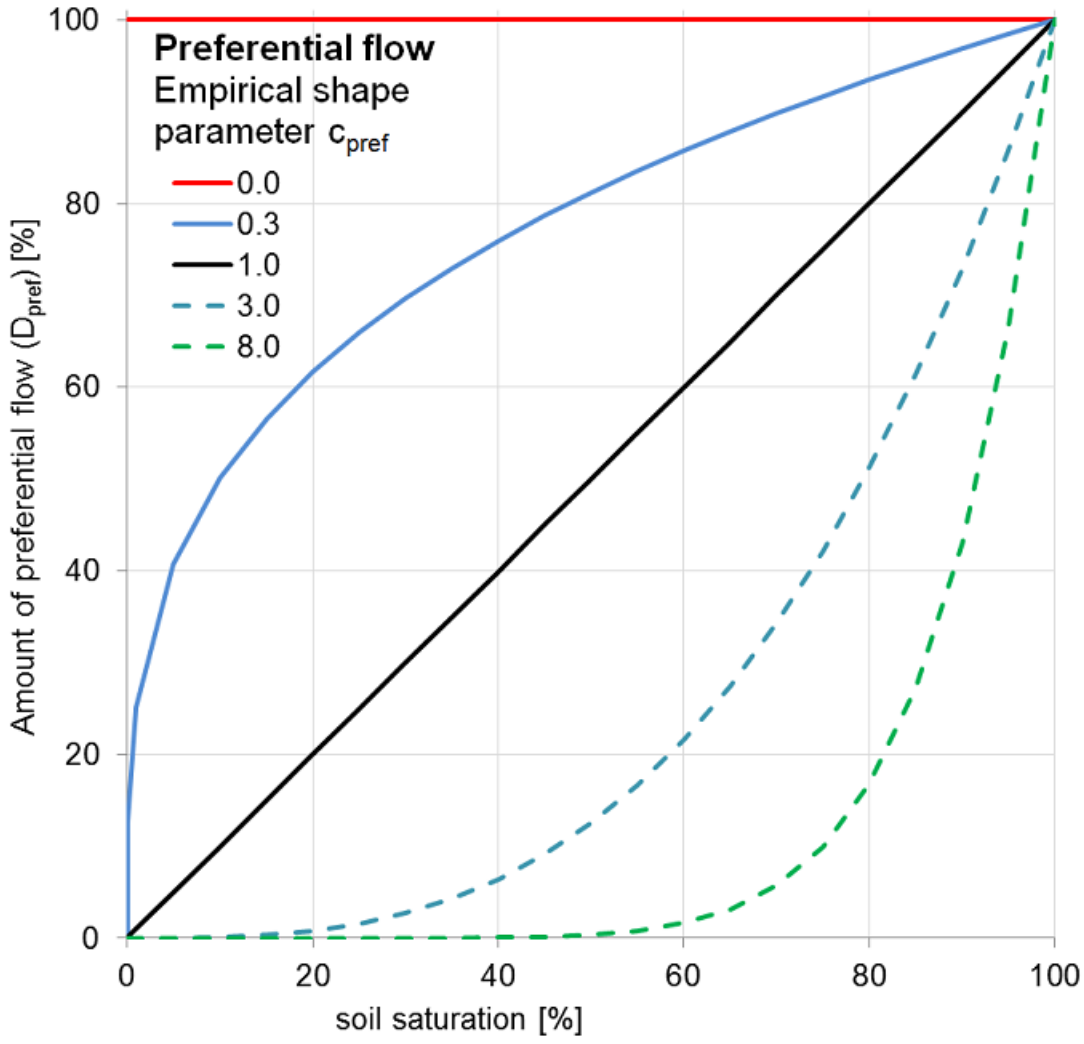


Figure 24:

Figure: *Soil moisture and preferential flow relation.*

Groundwater

Groundwater storage and transport are modelled using two parallel linear reservoirs, similar to the approach used in the HBV-96 model (Lindström et al., 1997). The upper zone represents a quick runoff component, which includes fast groundwater and subsurface flow through macro-pores in the soil. The lower zone represents the slow groundwater component that generates the base flow. The outflow from the upper zone to the channel, $Q_{uz}[mm]$ equals:

$$Q_{uz} = \frac{1}{T_{uz}} \cdot UZ \cdot \Delta t$$

where T_{uz} is a reservoir constant $[days]$ and UZ is the amount of water that is stored in the upper zone $[mm]$. Similarly, the outflow from the lower zone is given by:

$$Q_{lz} = \frac{1}{T_{lz}} \cdot LZ \cdot \Delta t$$

Here, T_{lz} is again a reservoir constant $[days]$, and LZ is the amount of water that is stored in the lower zone $[mm]$. The values of both T_{uz} and T_{lz} are obtained by calibration. The upper zone also provides the inflow into the lower zone. For each time step, a fixed amount of water percolates from the upper to the lower zone:

$$D_{uz,lz} = min(GW_{perc} \cdot \Delta t, UZ)$$

Here, GW_{perc} $[\frac{mm}{day}]$ is a user-defined value that can be used as a calibration constant. For many catchments it is quite reasonable to treat the lower groundwater zone as a system with a closed lower boundary (i.e. water is either stored, or added to the channel). However, in some cases the closed boundary assumption makes it impossible to obtain realistic simulations. Because of this, it is possible to percolate a fixed amount of water out of the lower zone, as a loss D_{loss}:

$$D_{loss} = min(f_{loss} \cdot \Delta t, LZ)$$

In the previous version of LISFLOOD D_{loss}, was calculated as a fixed fraction of Q_{lz}, but this leads to a high dependency of D_{loss} from GW_{perc} and LZ. For example if either GW_{perc} or LZ is quite low the parameter D_{loss} turns out to be meaningless.

The loss fraction, f_{loss}

–

, equals 0 for a completely closed lower boundary. If f_{loss} is set to 1, all outflow from the lower zone is treated as a loss. Water that flows out of the lower zone through D_{loss} is quite literally 'lost' forever. Physically, the loss term could represent water that is either lost to deep groundwater systems (that do not necessarily follow catchment boundaries), or even groundwater extraction wells. When using the model, it is suggested to use f_{loss} with some care; start with a value of zero, and only use any other value if it is impossible to get satisfactory results by adjusting the other calibration parameters. At each time step, the amounts of water in the upper and lower zone are updated for the in- and outgoing fluxes, i.e.:

$$UZ_t = UZ_{t-1} + D_{2,gw} - D_{uz,lz} - Q_{uz}$$

$$LZ_t = LZ_{t-1} + D_{uz,lz} - Q_{lz} - D_{loss}$$

Note that these equations are again valid for the permeable fraction of the pixel only: storage in the direct runoff fraction equals 0 for both UZ and LZ.

Dynamic wave option

Introduction

This page describes the LISFLOOD dynamic wave routine, and how it is used. A straightforward iteration using an Euler solution scheme is used to solve these equations. Dynamic wave routing is *optional*, and can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="dynamicWave" choice="1" />
```

Is this paragraph still up to date?

Note: The current implementation of the dynamic wave function in PCRaster is not a complete dynamic wave formulation according to the summary of the Saint Venant equations as discussed in Chow (1988). The implementation currently consists of the friction force term, the gravity force term and the pressure force term and should therefore be correctly characterised as a diffusion wave formulation. The equations are solved as an explicit, finite forward difference scheme.

Time step selection

The current dynamic wave implementation requires that all equations are solved using a time step that is much smaller (order of magnitude: seconds-minutes) than the typical overall time step used by LISFLOOD (order of magnitude: hours-day). More specifically, during one (sub) time step no water should be allowed to travel more than 1 cell downstream, i.e.:

$$\Delta' t_{dyn} \leq \frac{\Delta x}{V + c_d}$$

where $\Delta' t_{dyn}$ is the sub-step for the dynamic wave $[seconds]$, x is the length of one channel element (pixel) $[m]$, V is the flow velocity $[\frac{m}{s}]$ and c_d is dynamic wave celerity $[\frac{m}{s}]$.

The dynamic wave celerity can be calculated as (Chow, 1988):

$$c_d = \sqrt{gy}$$

where g is the acceleration by gravity $[\frac{m}{s^2}]$ and y is the depth of flow $[m]$. For a cross-section of a regular geometric shape, y can be calculated from the channel dimensions. Since the current dynamic wave routine uses irregularly shaped cross-section data, we simply assume than y equals the water level above the channel bed.

The flow velocity is simply:

$$V = \frac{Q_{ch}}{A}$$

where Q_{ch} is the discharge in the channel $[\frac{m^3}{s}]$, and A the cross-sectional area $[m^2]$.

The Courant number for the dynamic wave, C_{dyn}, can now be computed as:

$$C_{dyn} = \frac{(V + c_d)\Delta t}{\Delta x}$$

where t is the overall model time step

$$s$$

.

The number of sub-steps is then given by:

$$SubSteps = \max(1, roundup(\frac{C_{dyn}}{C_{dyn,crit}}))$$

where $C_{dyn,crit}$ is the critical Courant number. The maximum value of the critical Courant number is 1; in practice it is safer to use a somewhat smaller value (although if you make it too small the model becomes excessively slow). It is recommended to stick to the default value (0.4) that is used the settings file template.

Input data

A number of additional input files are necessary to use the dynamic wave option. First, the channel stretches for which the dynamic wave is to be used are defined on a Boolean map. Next, a cross-section identifier map is needed that links the (dynamic wave) channel pixels to the cross-section table (see further down). A channel bottom level map describes the bottom level of the channel (relative to sea level). Finally, a cross-section table describes the relationship between water height (H), channel cross-sectional area (A) and wetted perimeter (P) for a succession of H levels.

The following table lists all required input:

Table: *Input requirements dynamic wave routine*

Maps	Default name	Description	Units	Remarks
ChannelsDynamic	chandyn.map	dynamic wave channels (1,0)	-	Boolean
ChanCrossSections	chanxsect.map	channel cross section IDs	-	nominal
ChanBottomLevel	chblevel.map	channel bottom level	m	
Tables				
TabCrossSections	chanxsect.txt	cross section parameter table (H,A,P)	H: m A: m^2 P: m	

Layout of the cross-section parameter table

The cross-section parameter table is a text file that contains –for each cross-section- a sequence of water levels (H) with their corresponding cross-sectional area (A) and wetted perimeter (P). The format of each line is as follows:

ID H A P

As an example:

```
+--------------------------------+
| ID H A P                       |
| 167 0 0 0                      |
| 167 1.507 103.044 114.183      |
| 167 3.015 362.28 302.652       |
| 167 4.522 902.288 448.206      |
| 167 6.03 1709.097 600.382      |
| 167 6.217 1821.849 609.433     |
| 167 6.591 2049.726 615.835     |
| 167 6.778 2164.351 618.012     |
| 167 6.965 2279.355 620.14      |
| 167 7.152 2395.037 626.183     |
| 167 7.526 2629.098 631.759     |
| 167 7.713 2746.569 634.07      |
| 167 7.9 2864.589 636.93        |
| 167 307.9 192201.4874 5225.1652 |
+--------------------------------+
```

Note here that the first H-level is always 0, for which A and P are (of course) 0 as well. For the last line for each cross-section it is recommended to use some very (i.e. unrealistically) high H-level. The reason for doing this is that the dynamic wave routine will crash if during a simulation a water level (or cross-sectional area) is simulated which is beyond the range of the table. This can occur due to a number of reasons (e.g. if the measured cross-section is incomplete, or during calibration of the model). To estimate the corresponding values of A and P one could for example calculate dA/dH and dP/dH over the last two 'real' (i.e. measured) H-levels, and extrapolate the results to a very high H-level.

The number of H/A/P combinations that are used for each cross section is user-defined. LISFLOOD automatically interpolates in between the table values.

Using the dynamic wave

The 'lfuser' element contains two parameters that can be set by the user: *CourantDynamicCrit* (which should always be smaller than '1' and a parameter called *DynWaveConstantHeadBoundary*, which defines the boundary condition at the most downstream cell. All remaining dynamic-wave related input is defined in the 'lfbinding' element, and doesn't require any changes from the user (provided that all default names are used, all maps are in the standard 'maps' directory and the profile table is in the 'tables' directory). In 'lfuser' this will look like this:

```
<comment>
************************************************************
DYNAMIC WAVE OPTION
************************************************************
</comment>
<textvar name="CourantDynamicCrit" value="0.5">
<comment>
Critical Courant number for dynamic wave
value between 0-1 (smaller values result in greater numerical
accuracy,
but also increase computational time)
</comment>
</textvar>
<textvar name="DynWaveConstantHeadBoundary" value="0">
<comment>
Constant head [m] at most downstream pixel (relative to altitude
at most downstream pixel)
</comment>
</textvar>
```

Inflow hydrograph option

Introduction

This page describes the LISFLOOD inflow hydrograph routine, and how it is used. Inflow hydrographs are *optional*, and can be activated by adding the following line to the 'lfoptions' element in the LISFLOOD settings file:

```
<setoption name="inflow" choice="1" />
```

Description of the inflow hydrograph routine

When using the inflow hydrograph option, time series of discharge $[\frac{m^3}{s}]$ are added at some user-defined location(s) on the channel network. The inflow is added as side-flow in the channel routing equations (this works for both kinematic and dynamic wave). *Negative* inflows (i.e. outflows) are also possible, but large outflow rates may sometimes result in numerical problems in the routing equations. If you use a negative inflow rate, we advise to carefully inspect the model output for any signs of numerical problems (i.e. strange oscillations in simulated discharge, generation of missing values). Also check the mass balance time series after your simulation (numerical problems often result in unusually large mass balance errors).

Using inflow hydrographs

The table below lists the input requirements for the inflow hydrograph option. All you need is a map that defines where you want to add the inflow, and a time series with the corresponding inflow rates.

Table: *Input requirements inflow hydrograph routine.*

Maps	Default name	Description	Units	Remarks
InflowPoints	-	locations for inflow hydrographs	-	nominal
Time series				
QInTS	-	inflow hydrograph(s)	$\frac{m^3}{s}$	

Using the inflow hydrograph option involves **four steps**:

1) Create a (nominal) PCRaster map with unique identifiers that point to the location(s) where you want to insert the inflow hydrograph(s)

2) Save the inflow hydrograph(s) in PCRaster time series format; inflow hydrographs need to be given in $[\frac{m^3}{s}]$

IMPORTANT: PCRaster assumes that the first data series in the time series file (i.e. the second column, since the first column contains the time step number) corresponds to unique identifier 1 on the InflowPoints map; the second series to unique identifier 2, and so on. So, even if your InflowPoints map only contains (as an example) identifiers 3 and 4, you still need to include the columns for identifiers 1 and 2!! The best thing to do in such a case is to fill any unused columns with zeroes (0). Also, your inflow hydrograph time series should always start at t=1, even if you set StepStart to some higher value. For more info on time series files please have a look at the PCRaster documentation.

3) Make sure that the names of the map and time series are defined in the settings file

In the 'lfuser' element (replace the file paths/names by the ones you want to use):

```
<group>
<comment>
************************************************************
INFLOW HYDROGRAPH (OPTIONAL)
************************************************************
</comment>
<textvar name="InflowPoints"
value="/floods2/yourhomedir/yourcatchment/maps/inlets.map">
<comment>
OPTIONAL: nominal map with locations of (measured)
inflow hydrographs [cu m / s]
</comment>
</textvar>
<textvar name="QInTS"
value="/floods2/yourhomedir/yourcatchment/inflow/inflow.tss">
<comment>
OPTIONAL: observed or simulated input hydrographs as
time series [cu m / s]
Note that identifiers in time series correspond to
InflowPoints map (also optional)
</comment>
</textvar>
</group>
```

4) Activate the inflow hydrograph option by adding the following line to the 'lfoptions' element:

```
<setoption name="inflow" choice="1" />
```

Now you are ready to run the model with the inflow hydrograph.

Substituting subcatchments with measured inflow hydrographs

One of the most common uses of the inflow hydrograph option is this: suppose we have a catchment where we only want to simulate the downstream part. If measured time series of discharge are available for the upstream catchment(s), we can use these to represent the inflow into the more downstream part. The Figure below shows an example, where we have measured discharge of subcatchment A (just before it enters the main river).

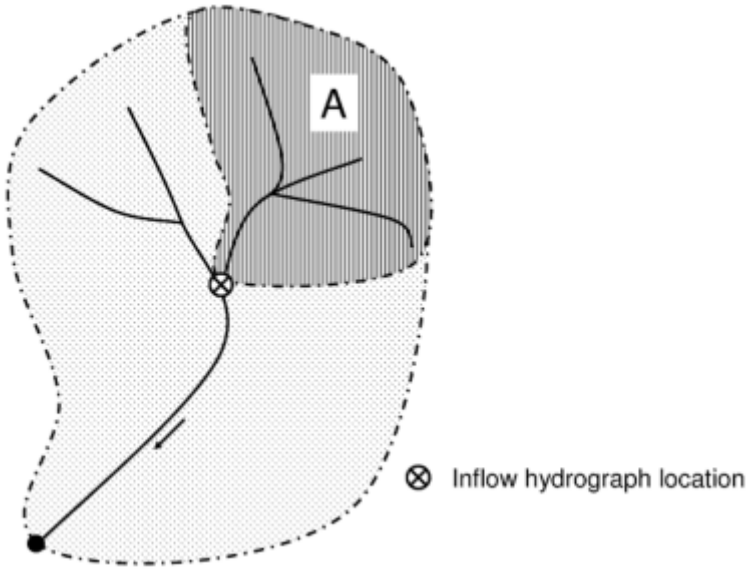


Figure 25:

Figure: *Using the inflow hydrograph using measured discharge of subcatchment A. MaskMap must have boolean(0) (or missing value) for subcatchment A, see text below for explanation.*

In this case it is important to pay special attention to two issues:

1) Exclude subcatchments from MaskMap

First, make sure that subcatchment *A* is *excluded* (i.e. have boolean(0) or missing value) on LISFLOOD's *MaskMap* (which defines which pixels are included in the calculations and which are not). If you include it, LISFLOOD will first *simulate* discharge coming out of subcatchment *A*, and then *add* the (measured) inflow on top of it! Of course this doesn't make any sense, so always be careful which areas are included in your simulation and which are not.

2) Make sure your inflow points are where you need them

If you already have all gauge locations on a map, these mostly cannot be used directly as inflow hydrograph locations. The reason is simple: suppose –in our previous example– we know the outflow point of subcatchment *A*. This point is the most downstream point within that subcatchment. However, the flow out of subcatchment *A* is actually added to the main river one cell downstream! Also, if we exclude subcatchment *A* from our simulation (as explained in the foregoing), this means we also exclude the outflow point of that subcatchment. Because of this, *inflow* points into the main river are usually located one pixel downstream of the *outflow* points of the corresponding subcatchment. If you already have a (nominal) map of of your subcatchments, a PCRaster script exists Script needs uploadin gin the LISFLOOD utilities repository and link needs to be added that automatically calculates the corresponding out- and inflow points.

Double kinematic wave option

Introduction

This annex describes the LISFLOOD double kinematic wave routine, and how it is used. Double kinematic wave routing is *optional*, and can be activated by adding the following line to the 'lfoptions' element the to LISFLOOD settings file (add link):

```
<setoption name="SplitRouting" choice="1" />
```

Background

The flow routing is done by the kinematic wave approach. Therefore two equations have to be solved:

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = q\rho g A(S_0 - S_f) = 0$$

where $A = \alpha \cdot Q^{\beta}$

The continuity equation momentum equation as expressed by Chow et al. 1988. With decreasing inflow the peaks of the resulting outflow will be later in time (see Figure below for a simple kinematic wave calculation). The wave propagation slows down because of more friction on the boundaries.

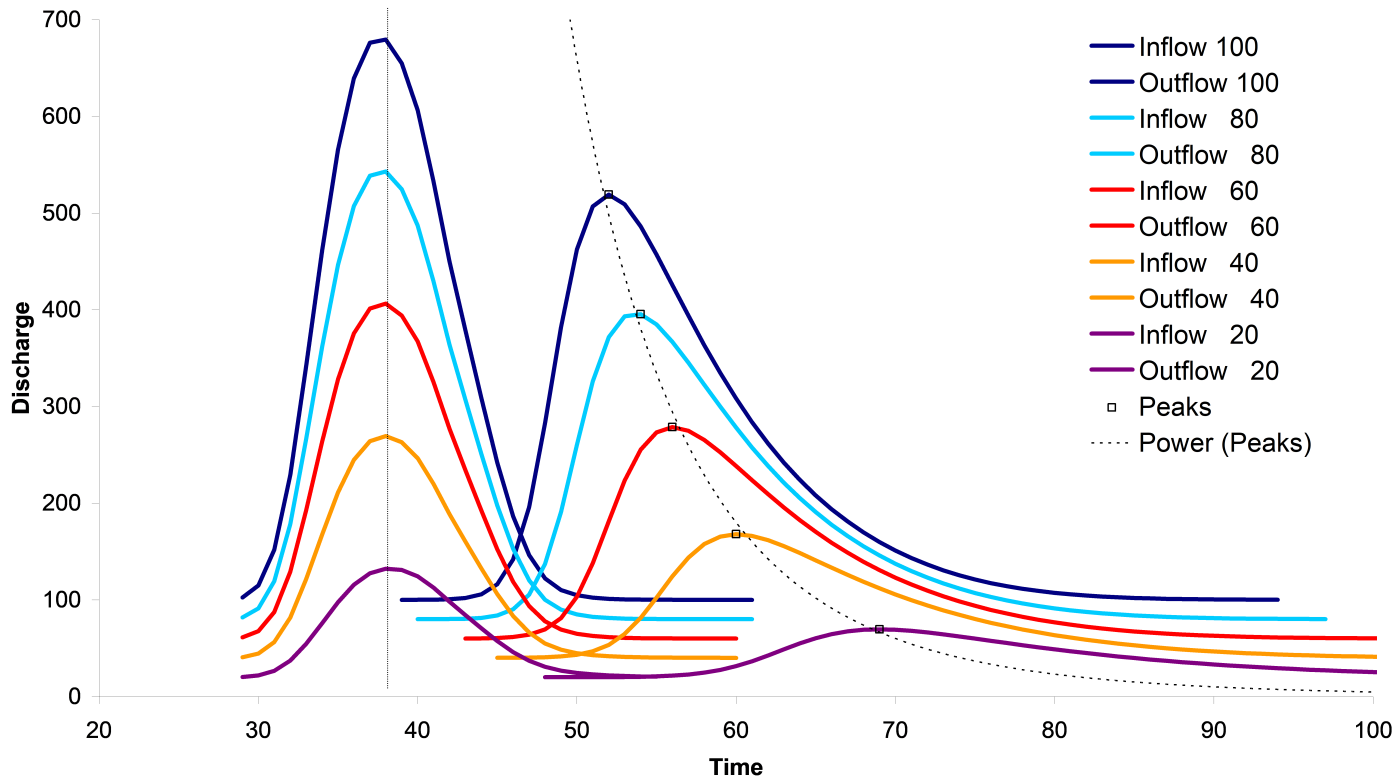


Figure 26:

Figure: *Simulated outflow for different amount of inflow wave propagation gets slower.*

This is realistic if your channel looks like this:

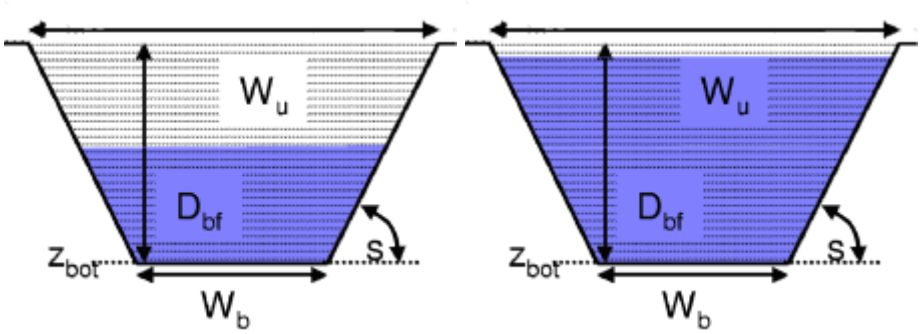


Figure: *Schematic cross section of a channel with different water level.*

But a natural channel looks more like this:

Figure: *Schematic cross section of a natural channel.*

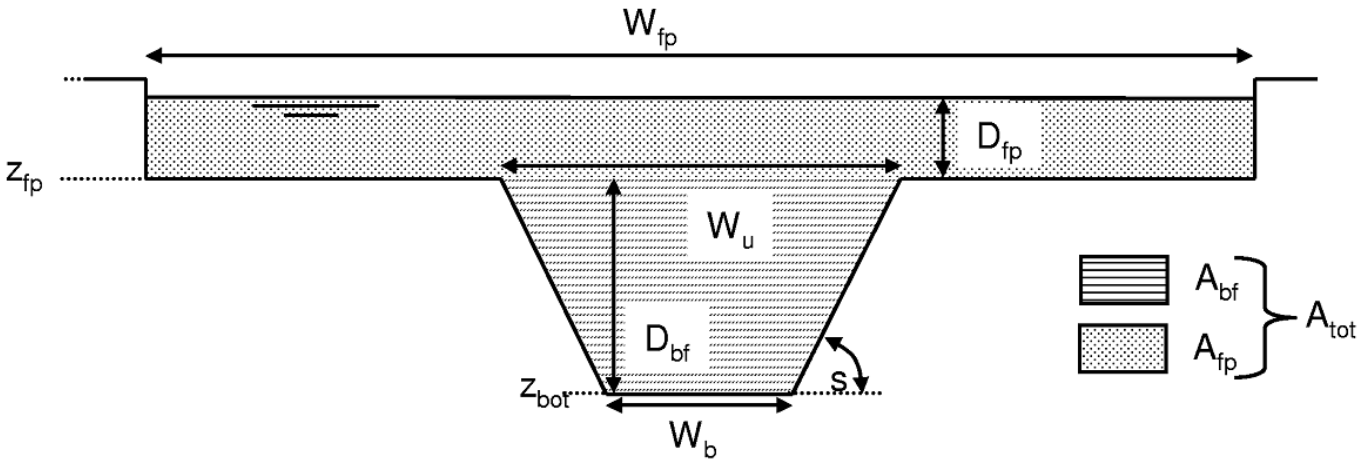


Figure 27:

Which means, opposite to the kinematic wave theory, the wave propagation gets slower as the discharge is increasing, because friction is going up on floodplains with shrubs, trees, bridges. Some of the water is even stored in the floodplains (e.g. retention areas, seepage retention). As a result of this, a single kinematic wave cannot cover these different characteristics of floods and floodplains.

Double kinematic wave approach

The double kinematic approach splits up the channel in two parts (see figure below):

1. bankful routing
2. over bankful routing

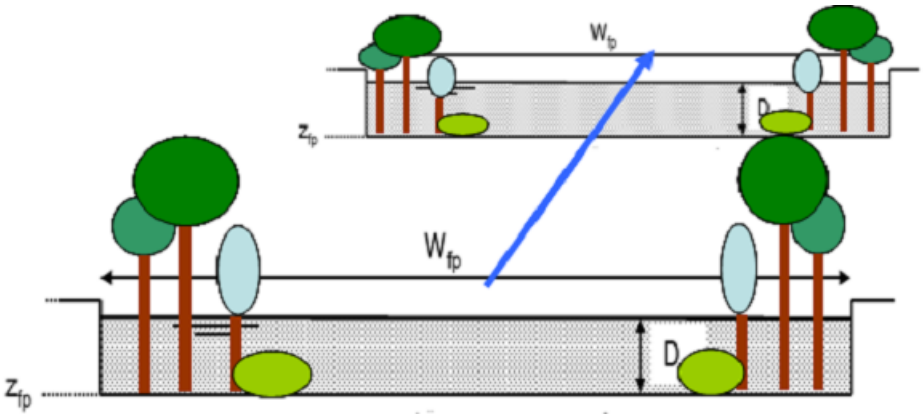


Figure 28:

Figure: *Channel is split in a bankful and over bankful routing*

Similar methods are used since the 1970s e.g. as multiple linear or non linear storage cascade (Chow, 1988). The former forecasting model for the River Elbe (ELBA) used a three stages approach depending on discharge (Fröhlich, 1996).

Using double kinematic wave

No additional maps or tables are needed for initializing the double kinematic wave. A normal run ('InitLisflood'=0) requires an additional map derived from the prerun ('InitLisflood'=1). A 'warm' start add link (using initial values from a previous run) requires two additional maps with state variables for the second (over 'bankful' routing).

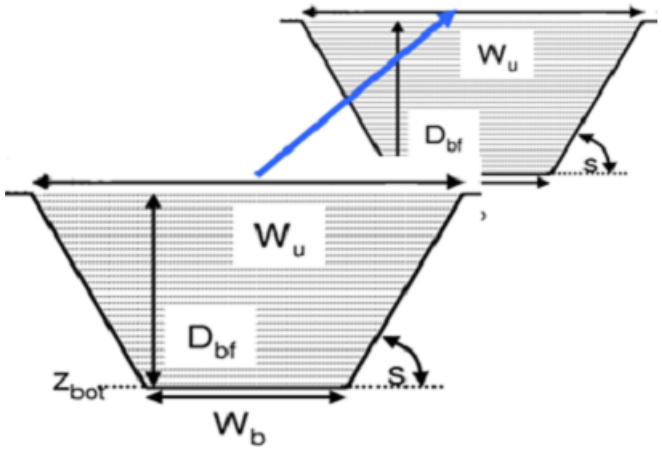


Figure 29:

Table: *Input/output double kinematic wave.*

Maps	Default name	Description	Units	Remarks
Average discharge	avgdis.map	Average discharge	$\frac{m^3}{s}$	Produced by prerun
CrossSection2AreaInitValue	chanbnkfrac00.xxx	channel crosssection for 2nd routing channel	m^2	Produced by option 'repStateMaps' or 'repEndMaps'
PrevSideflowInitValue	chside00.xxx	sideflow into the channel	mm	

Using the double kinematic wave approach option involves **three steps**:

1) In the 'lfuser' element (replace the file paths/names by the ones you want to use):

```
</textvar>
<textvar name="CalChanMan2" value="8.5">
<comment>
Channel Manning's n for second line of routing
</comment>
</textvar>
<textvar name="QSplitMult" value="2.0">
<comment>
Multiplier applied to average Q to split into a second line of routing
</comment>
</textvar>
```

CalChanMan2 is a multiplier that is applied to the Manning's roughness maps of the over bankful routing [-]

QSplitMult is a factor to the average discharge to determine the bankful discharge. The average discharge map is produced in the initial run (the initial run is already needed to get the groundwater storage). Standard is set to 2.0 (assuming over bankful discharge starts at 2.0 · average discharge).

2) Activate the double kinematic wave option by adding the following line to the 'lfoptions' element:

```
<setoption name="SplitRouting" choice="1" />
```

3) Run LISFLOOD first with

```
<setoption name="InitLisflood" choice="1" />
```

and it will produce a map of average discharge [$\frac{m^3}{s}$] in the initial folder. This map is used together with the QSplitMult factor to set the value for the second line of routing to start.

For a 'warm start' these initial values are needed

```
<textvar name="CrossSection2AreaInitValue" value="-9999">
<comment>
initial channel crosssection for 2nd routing channel
-9999: use 0
</comment>
```

Automatic change of the number of sub steps (optional)

For the new method the kinematic wave has to be applied two times.

The calculation of kinematic wave is the most time consuming part in a LISFLOOD run (in general but also depending on the catchment). The use of the double kinematic wave makes it necessary to calculate the kinematic wave two times and increasing the computing time. To counteract this, an option is put in place to change automatically the number of sub steps for channel routing.

Double kinematic wave routing is *optional*, and can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="VarNoSubStepChannel" choice="1" />
```

This will calculate the number of sub steps for the kinematic wave according to the discharge. Less number of steps for low and average flow condition, more sub steps for flooding condition because the higher velocity of water.

Activating this option needs to be done before the prerun ('InitLisflood'=1) because the maximum celerity of wave propagation (chanckinmax.map) is created as another initial map and used in the 'normal' runs.

The minimum and maximum number of sub steps can be set in the settings file:

```
<comment>
************************************************************
Variable Channel NoSubStepChannel
************************************************************
</comment>
<textvar name="UpLimit" value="1.0E+9">
<comment>
Upstream area do be included in max. celerity
</comment>
</textvar>
<textvar name="MinNoStep" value="5">
<comment>
minimum number of sub steps for channel
</comment>
</textvar>
<textvar name="ChanA" value="30">
<comment>
max. NoSubStepsChannel = ChanA-ChanB
</comment>
</textvar>
<textvar name="ChanB" value="10">
<comment>
For calculating the min. No. of substeps
</comment>
</textvar>
```

UpLimit is the minimum upstream area do be included in the calculation of the maximum celerity of wave propagation [m^2]

MinNoStep is the absolute minimum number of sub steps for channel routing [-]

ChanA for calculating the maximum number of sub steps for channel routing [-]

ChanB for calculating the minimum number of sub steps for channelrouting [-]

Simulation of lakes

Introduction

This pages describes the LISFLOOD lake routine, and how it is used. The simulation of lakes is *optional*, and it can be activated by adding the following line to the 'lfoptions' element in the LISFLOOD settings file:

```
<setoption name="simulateLakes" choice="1" />
```

Lakes can be simulated on channel pixels where kinematic wave routing is used. The routine does *not* work for channel stretches where the dynamic wave is used!

Description of the lake routine

Lakes are simulated as points in the channel network. The Figure below shows all computed in- and outgoing fluxes. Lake inflow equals the channel flow upstream of the lake location. The flow out of the lake is computed using the following rating curve (e.g. Maidment, 1992):

$$O_{lake} = A(H - H_0)^B$$

with: O_{lake}: Lake outflow rate $[\frac{m^3}{s}]$ H: Water level in lake $[m]$ H_0: Water level at which lake outflow is zero $[m]$ A, B: Constants

–

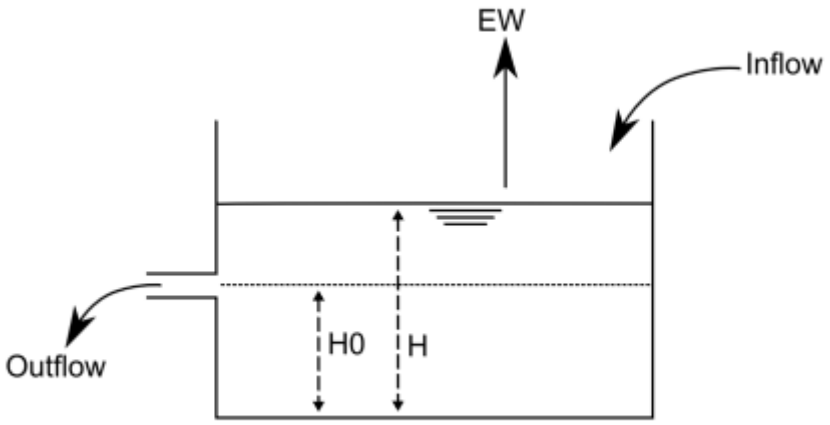


Figure 30:

Figure: *Schematic overview of the simulation of lakes. H_0 is the water level at which the outflow is zero; H is the water level in the lake and EW is the evaporation from the lake*

Both H and H_0 can be defined relative to an arbitrary reference level. Since the outflow is a function of the *difference* between both levels, the actual value of this reference level doesn't matter if $H > H_0$. However, it is advised to define both H and H_0 relative to the *average bottom level* of the lake. This will result in more realistic simulations during severe drought spells, when the water level drops below H_0 (in which case lake outflow ceases). The value of constant A can be approximated by the width of the lake outlet in meters, and B is within the range 1.5-2 (reference?). Lake evaporation occurs at the potential evaporation rate of an open water surface.

Initialisation of the lake routine

Because lakes (especially large ones) tend to produce a relatively slow response over time, it is important to make sure that the initial lake level is set to a more or less sensible value. Just as is the case with the initialisation of the lower groundwater zone, LISFLOOD has a special option that will compute a steady-state lake level and use this as the initial value. The steady-state level is computed from the water balance of the lake. If V_l is the total lake volume $[m^3]$, the rate of change of V_l at any moment is given by the continuity equation:

$$\frac{dV_l}{dt} = I(t) - O(t)$$

where I and O are the in- and outflow rates, respectively. For a steady-state situation the storage remains constant, so:

$$\frac{dV_l}{dt} = 0 \quad \Leftrightarrow \quad I(t) - O(t) = 0$$

Substituting all in- and outflow terms gives:

$$I_l - EW_l - A(H - H_0)^B = 0$$

where I_l is the inflow into the lake and EW_l the lake evaporation (both expressed in $\frac{m^3}{s}$). Re-arranging gives the steady-state lake level:

$$H_{ss} = H_0 + (\frac{I_l - EW_l}{A})^{\frac{1}{B}}$$

LISFLOOD calculates the steady-state lake level based on a user-defined average net inflow ($= I_l - EW_l$). The average net inflow can be estimated using measured discharge and evaporation records. If measured discharge is available just *downstream* of the lake (i.e. the *outflow*), the (long-term) average outflow can be used as the net inflow estimate (since, for a steady state situation, inflow equals outflow). If only inflow is available, all average inflows should be summed, and the average lake evaporation should be subtracted from this figure.

Here a worked example. Be aware that the calculation can be less straightforward for very large lakes with multiple inlets (which are not well represented by the current point approach anyway):

EXAMPLE: Calculation of average net lake inflow

Lake characteristics - lake area: $215 \cdot 10^6\ m^2$
- mean annual discharge downstream of lake: $293\ \frac{m^3}{s}$
- mean annual discharge upstream of lake: $300\ \frac{m^3}{s}$
- mean annual evaporation: $1100\ \frac{mm}{yr}$

METHOD 1: USING AVERAGE OUTFLOW
Assuming lake is in quasi steady-state:
average net inflow = average net outflow = $293 \frac{m^3}{s}$

METHOD 2: USING AVERAGE INFLOW AND EVAPORATION
Only use this method if no outflow data are available

1. Express lake evaporation in $m^3 s^{-1}$:

   $\frac{1100 \frac{mm}{yr}}{1000} = 1.1\ \frac{m}{yr}$

   $1.1\ \frac{m}{yr} \cdot 215 \cdot 10^6 m^2 = 2.37 \cdot 10^8 \frac{m^3}{yr}$

   $\frac{2.37 \cdot 10^8 \frac{m^3}{yr}}{365\ days\ \cdot\ 86400 seconds} = 7.5 \frac{m^3}{s}$
2. Compute net inflow:
   net inflow = $300 \frac{m^3}{s}\ - 7.5\ \frac{m^3}{s} = 292.5\ \frac{m^3}{s}$

Preparation of input data

The lake locations are defined on a (nominal) map called '*lakes.nc*'. It is important that all lakes are located on a channel pixel (you can verify this by displaying the lake map on top of the channel map). Also, since each lake receives its inflow from its upstream neighbouring channel pixel, you may want to check if each lake has any upstream channel pixels at all (if not, the lake will just gradually empty during a model run!). The lake characteristics are described by 4 tables. The following Table lists all required input:

Table: *Input requirements lake routine.*

Maps	Default name	Description	Units	Remarks
LakeSites	lakes.map	lake locations	-	nominal
Tables				
TabLakeArea	lakearea.txt	lake surface area	m^2	
TabLakeH0	lakeh0.txt	water level at which lake outflowis zero	m	relative to average lake bottom level
TabLakeA	lakea.txt	lake parameter A	-	outlet width in meters
TabLakeB	lakeb.txt	lake parameter B	-	1.5-2

Note: When you create the map with the lake locations, pay special attention to the following: if a lake is located on the most downstream cell (i.e. the outflow point, see Figure below), the lake routine may produce erroneous output. In particular, the mass balance errors cannot be calculated correctly in that case. The same applies if you simulate only a sub-catchment of a larger map (by selecting the subcatchment in the mask map). This situation can usually be avoided by extending the mask map by one cell in downstream direction.

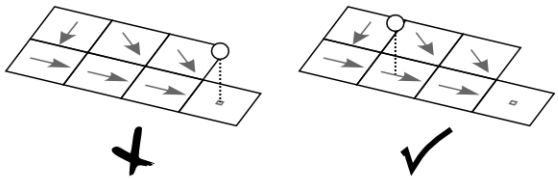


Figure 31:

Figure: *Placement of the lakes: lakes on the outflow point (left) result in erroneous behavior of the lake routine.*

Preparation of settings file

All in- and output files need to be defined in the settings file. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element. Just make sure that the map with the lake locations is in the "maps" directory, and all tables in the 'tables" directory. If this is the case, you only have to specify the initial lake level and –if you are using the steady-state option- the mean net lake inflow (make this a map if you're simulating multiple lakes simultaneously). Both can be set in the 'lfuser' element. *LakeInitialLevelValue* can be either a map or a single value. Setting *LakeInitialLevelValue* to *-9999* will cause LISFLOOD to calculate the steady-state level. So we add this to the 'lfuser' element (if it is not there already):

```
<group>
<comment>
************************************************************
LAKE OPTION
************************************************************
</comment>
<textvar name="LakeInitialLevelValue" value="-9999">
<comment>
```

```
Initial lake level [m]
-9999 sets initial value to steady-state level
</comment>
</textvar>
<textvar name="LakeAvNetInflowEstimate" value="292.5">
<comment>
Estimate of average net inflow into lake (=inflow -- evaporation)
[cu m / s]
Used to calculated steady-state lake level in case
LakeInitialLevelValue
is set to -9999
</comment>
</textvar>
</group>
```

Finally, you have to tell LISFLOOD that you want to simulate lakes! To do this, add the following statement to the 'lfoptions' element:

```
<setoption name="simulateLakes" choice="1" />
```

Now you are ready to run the model. If you want to compare the model results both with and without the inclusion of lakes, you can switch off the simulation of lakes either by:

- Removing the 'simulateLakes' statement from the 'lfoptions element, or
- changing it into <setoption name="simulateLakes" choice="0" />

Both have exactly the same effect. You don't need to change anything in either 'lfuser' or 'lfbinding'; all file definitions here are simply ignored during the execution of the model.

Lake output files

The lake routine produces 4 additional time series and one map (or stack), as listed in the following table:

Table: *Output of lake routine.*

Maps	Default name	Description	Units
LakeLevelState	lakhxxxx.xxx	lake level at last time step	m
Time series			
LakeInflowTS	qLakeIn.tss	inflow into lakes	$\frac{m^3}{s}$
LakeOutflowTS	qLakeOut.tss	flow out of lakes	$\frac{m^3}{s}$
LakeEWTS	EWLake.tss	lake evaporation	mm
LakeLevelTS	hLake.tss	lake level	m

Note that you can use the map with the lake level at the last time step to define the initial conditions of a succeeding simulation, e.g.:

```
<textvar name="LakeInitialLevelValue" value="/mycatchment/lakh0000.730">
```

Read and write NetCDF files

To read and write NetCDF files is *optional*, and has to be activated under the 'lfoptions' element in the LISFLOOD settings file.

Reading NetCDF files

LISFLOOD can read files containing forcing data and static maps both as NetCDF single map (without "time" variable) and as NetCDF stack (with "time" variable).

Reading of NetCDF files is activated using readNetcdfStack switch in lfoption section in Settings XML file:

```
<setoption choice="1"  name="readNetcdfStack"/>
```

If NetCDF file contains "time" variable, LISFLOOD reads NetCDF files by timestamps. Correspondence between LISFLOOD time steps and NetCDF timestamps is automatically computed within the model. LISFLOOD can run on any sub-period included in forcings data.

NetCDF forcings files (pr, ta, e0, es, et) are completely independent from LISFLOOD settings, meaning they can cover any period of time starting from any date, but they must include the entire LISFLOOD simulation period. A check is performed at the beginning of the simulation and an error message is provided if simulation period is outside forcings maps period. Missing maps are not allowed for forcings data and checks are in place to prevent using daily maps to perform sub-daily simulations.

Particular attention must be paid when running LISFLOOD using time steps. Time steps set in Settings XML file always refer to the date specified as CalendarDayStart. Time step values will be automatically converted to dates and corresponding date values will be read from NetCDF files.

Writing NetCDF files

LISFLOOD can write both NetCDF single maps (without "time" variable) and as NetCDF stacks (with "time" variable). Writing of NetCDF files is activated using switches in lfoption section in Settings XML file:

```
<setoption choice="1"  name="writeNetcdfStack"/>

<setoption choice="1"  name="writeNetcdf"/>
```

End files and State files can be saved in NetCDF file format, state files can be saved for a specified sub-period (sub-period can only be set using time steps) within the simulation period using:

```
<textvar name="ReportSteps" value="2801..9999">
```

Overview

The model description under 'STANDARD LISFLOOD PROCESSES' covers the processes that are simulated in a 'standard' LISFLOOD run. However, LISFLOOD holds a wide range of additional options of two types: 1) simulate additional features
2) write additional output.

Additional simulation options

Many additional options have been developed to **simulate** all kind of **additional features**, such as e.g.:

- Including: reservoirs, polder, lakes, inflow hydrographs and transmission losses
- chosing among different routing routines: double kinematic wave routing or dynamic wave routing
- Simulating water levels, water use and soil moisture

If you like to use an additional option you have to 'activate' it in the LISFLOOD settings file (**INSERT LINK TO SETTINGS FILE HERE**.) under the 'lfoptions' element. Each element under this option section represents a switch with "1" equal to "on", and "0" to "off". The table below shows all the currently implemented additional simulation options including their corresponding defaults. You can activate as many options as you want (or none at all) by setting the switch to 1. This way you can tell the model exactly which processes to calculate and which not.

Note that each option generally requires additional items in the settings file. For instance, using the inflow hydrograph option requires an input map and time series, which have to be specified in the settings file. The template settings file that is provided with LISFLOOD always contains file definitions for all optional output maps and time series.

Table: *LISFLOOD additional simulation options.*

Option	Description	Default
gridSizeUserDefined	Get grid size attributes (length, area) from user-defined maps (instead of using map location attributes directly)	0
simulateReservoirs	Simulate retention and hydropower reservoirs (kin. wave only)	0
simulateLakes	Simulate unregulated lakes (kin. wave only)	0
simulatePolders	Simulate flood protection polders (dyn. wave only)	0
inflow	Use inflow hydrographs	0
dynamicWave	Perform dynamic wave channel routing	0
simulateWaterLevels	Simulate water levels in channel	0
TransLoss	Simulate transmission loss	0
SplitRouting	Simulate double kinematic wave	0
VarNoSubStepChannel	Use variable number of sub step for channel routing	0
wateruse	Simulate water use	0

Additional output options

Besides the standard LISFLOOD output (which is discharge and soil moisture), the user has the option to receive all kind of additional output files. The table below lists all currently implemented output options and their corresponding defaults.

In the LISFLOOD settings file (**INSERT LINK TO SETTINGS FILE HERE**.) the 'lfoptions' element gives you additional control over what LISFLOOD is doing. As with the simulation options also the output options are implemented as switches with "1" corrisponding to "on" and "0" to "off". This way you can tell the model exactly which output files are reported and which ones aren't. You can activate as many options as you want (or none at all). Remember that each option generally requires additional items in the settings file. For instance, if you want to report discharge maps at each time step, you will first have to specify under which name they will be written. The template settings file that is provided with LISFLOOD always contains file definitions for all optional output maps and time series.

Actually the output option and so many others are still missing. We have to make the check that all options are described!.)

Table: *LISFLOOD additional reporting options.*

Option	Description	Default
OUTPUT, TIME SERIES		
repDischargeTs	Report timeseries of discharge at gauge locations	1
repWaterLevelTs	Report timeseries of water level at gauge locations14	0
repStateSites	Report timeseries of all intermediate state variables at 'sites'	0
repRateSites	Report timeseries of all intermediate rate variables at 'sites'	0
repMeteoUpsGauges	Report timeseries of meteorological input, averaged over contributing area of each gauging station	0
repStateUpsGauges	Report timeseries of model state variables, averaged over contributing area of each gauging station	0
repRateUpsGauges	Report timeseries of model rate variables, averaged over contributing area of each gauging station	0
OUTPUT, MASS BALANCE		
repMBTs	Report timeseries of absolute cumulative mass balance error	1
repMBMMTs	Report timeseries of cumulative mass balance error expressed as mm water slice	1
OUTPUT, MAPS,DISCHARGE		

Option	Description	Default
repDischargeMaps	Report maps of discharge (for each time step)	0
repWaterLevelMaps	Report maps of water level in channel (for each time step)	0
OUTPUT, MAPS, STATE VARIABLES (all, at selected time steps)		
repStateMaps	Report maps of model state variables (as defined by "ReportSteps" variable)	1
repEndMaps	Report maps of model state variables (at last time step)	0
OUTPUT, MAPS, STATE VARIABLES		
repDSLRMaps	Report maps of days since last rain (for each time step)	0
repFrostIndexMaps	Report maps of frost index (for each time step)	0
repWaterDepthMaps	Report maps of depth of water layer on soil surface (for each time step)	0
repSnowCoverMaps	Report maps of snow cover (for each time step)	0
repCumInterceptionMaps	Report maps of interception storage (for each time step)	0
repTheta1Maps	Report maps of soil moisture layer 1(for each time step)	0
repTheta2Maps	Report maps of soil moisture layer 2 (for each time step)	0
repUZMaps	Report maps of upper zone storage (for each time step)	0
repLZMaps	Report maps of lower zone storage (for each time step)	0
repChanCrossSectionMaps	Report maps of channel cross-sectional area (for each time step)	0
OUTPUT, MAPS, METEOROLOGICAL FORCING VARIABLES		
repPrecipitationMaps	Report maps of precipitation (for each time step)	0
repTavgMaps	Report maps of average temperature (for each time step)	0
repETRefMaps	Report maps of potential reference evapotranspiration (for each time step)	0
repESRefMaps	Report maps of potential soil evaporation (for each time step)	0
repEWRefMaps	Report maps of potential open water evaporation (for each time step)	0
OUTPUT, MAPS, RATE VARIABLES		
repRainMaps	Report maps of rain (excluding snow!) (for each time step)	0
repSnowMaps	Report maps of snow (for each time step)	0
repSnowMeltMaps	Report maps of snowmelt (for each time step)	0
repInterceptionMaps	Report maps of interception (for each time step)	0
repLeafDrainageMaps	Report maps of leaf drainage (for each time step)	0
repTaMaps	Report maps of actual transpiration (for each time step)	0
repESActMaps	Report maps of actual soil evaporation (for each time step)	0
repEWIntMaps	Report maps of actual evaporation of intercepted water (for each time step)	0
repInfiltrationMaps	Report maps of infiltration (for each time step)	0
repPrefFlowMaps	Report maps of preferential flow (for each time step)	0
repPercolationMaps	Report maps of percolation from upper to lower soil layer (for each time step)	0

Option	Description	Default
repSeepSubToGWMaps	Report maps of seepage from lower soil layer to ground water (for each time step)	0
repGwPercUZLZMaps	Report maps of percolation from upper to lower ground water zone (for each time step)	0
repGwLossMaps	Report maps of loss from lower ground water zone (for each time step)	0
repSurfaceRunoffMaps	Report maps of surface runoff (for each time step)	0
repUZOutflowMaps	Report maps of upper zone outflow (for each time step)	0
repLZOutflowMaps	Report maps of lower zone outflow (for each time step)	0
repTotalRunoffMaps	Report maps of total runoff (surface + upper + lower zone) (for each time step)	0
OUTPUT, MAPS (MISCELLANEOUS)		
repLZAvInflowMap	Report computed average inflow rate into lower zone (map, at last time step)	0
repLZAvInflowSites	Report computed average inflow rate into lower zone (time series, at points defined on sites map)	0
repLZAvInflowUpsGauges	Report computed average inflow rate into lower zone (time series, averaged over upstream area of each gauge location)	0

Polder option

Introduction

This page describes the LISFLOOD polder routine, and how it is used. The simulation of polders is *optional*, and it can be activated by adding the following line to the 'lfoptions' element of the settings file add link to settings file here.:

```
<setoption name="simulatePolders" choice="1" />
```

Polders can be simulated on channel pixels where dynamic wave routing is used. The routine does *not* work for channel stretches where the kinematic wave is used!

Description of the polder routine

Polders are simulated as points in the channel network. The polder routine is adapted from Förster et. al (2004), and based on the weir equation of Poleni (Bollrich & Preißler, 1992). The flow rates from the channel to the polder area and vice versa are calculated by balancing out the water levels in the channel and in the polder, as shown in the following Figure:

Figure: *Schematic overview of the simulation of polders. p_b is the polder bottom level (above the channel bottom); w_c is the water level in the channel; h_c and h_p are the water levels above the polder in- / outflow, respectively*

From the Figure, it is easy to see that there can be three situations:

1. $h_c > h_p$: water flows out of the channel, into the polder. The flow rate, $q_{c,p}$, is calculated using:

$$\left| \begin{array}{l} q_{c,p} = \mu \cdot c \cdot b \cdot \sqrt{2g} \cdot h_c^{3/2} \\ c = \sqrt{1 - [\frac{h_p}{h_c}]^{16}} \end{array} \right.$$

   where b is the outflow width $[m]$, g is the acceleration due to gravity (9.81 $\frac{m}{s^2}$) and μ is a weir constant which has a value of 0.49. Furthermore is $q_{c,p}$ in $\frac{m}{s}$.

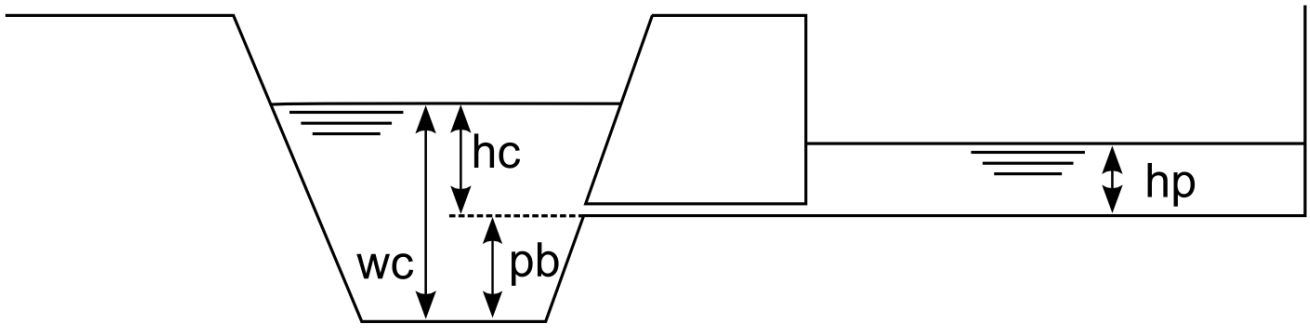


Figure 32:

2. $h_c < h_p$: water flows out of the polder back into the channel. The flow rate, $q_{p,c}$ is now calculated using:

$$\left| \begin{array}{l} q_{p,c} = \mu \cdot c \cdot b\sqrt{2g} \cdot h_p^{3/2} \\ c = \sqrt{1 - [\frac{h_c}{h_p}]^{16}} \end{array} \right.$$

3. $h_c = h_p$: no water flowing into either direction (note here that the minimum value of h_c is zero). In this case both $q_{c,p}$ and $q_{p,c}$ are zero.

Regulated and unregulated polders

The above equations are valid for *unregulated* polders. It is also possible to simulated *regulated* polders, which is illustrated in following Figure.

Figure: *Simulation of a regulated polder. Polder is closed (inactive) until user-defined opening time, after which it fills up to its capacity (flow rate according to Eq XXXX). Water stays in polder until user-defined release time, after which water is released back to the channel (flow rate according to Eq XXXX).*

Regulated polders are opened at a user-defined time (typically during the rising limb of a flood peak). The polder closes automatically once it is full. Subsequently, the polder is opened again to release the stored water back into the channel, which also occurs at a user-defined time. The opening- and release times for each polder are defined in two lookup tables (see Table below). In order to simulate the polders in *unregulated* mode these times should both be set to a bogus value of -9999. *Only* if *both* opening- and release time are set to some other value, LISFLOOD will simulate a polder in regulated mode. Since LISFLOOD only supports *one* single regulated open-close-release cycle per simulation, you should use regulated mode *only* for single flood events. For continuous simulations (e.g. long-tem waterbalance runs) you should only run the polders in unregulated mode.

Preparation of input data

The locations of the polders are defined on a (nominal) map called '*polders.map*'. Any polders that are *not* on a channel pixel are ignored by LISFLOOD, so you may want to check the polder locations before running the model (you can do this by displaying the polder map on top of the channel map). The current implementation of the polder routine may result in numerical instabilities for kinematic wave pixels, so for the moment it is recommended to define polders *only* on channels where the dynamic wave is used. Furthermore, the properties of each polder are described using a number of tables. All required input is listed in the following table:

Table: *Input requirements polder routine.*

Maps	**Defaultname**	**Description**	**Units**	**Remarks**
PolderSites	polders.map	polder locations	-	nominal

Maps	Defaultname	Description	Units	Remarks
Tables	**Defaultname**	**Description**	**Units**	**Remarks**
TabPolderArea	poldarea.txt	polder area	m^2	
TabPolderOFWidth	poldofw.txt	polder in- and outflow width	m	
TabPolderTotalCapacity	poldcap.txt	polder storage capacity	m^3	
TabPolderBottomLevel	poldblevel.txt	Bottom level of polder, measured from channel bottom level (see also Figure above)	m	
TabPolderOpeningTime	poldtopen.txt	Time at which polder is opened	*timestep*	
TabPolderReleaseTime	poldtrelease.txt	Time at which water stored in polder is released again	*timestep*	

Note that the polder opening- and release times are both defined a *time step* numbers (*not* days or hours!!). For *unregulated* polders, set both parameters to a bogus value of -9999, i.e.:

Is that still the same or is it now possible to do it with a date stamp?

```
10 -9999
15 -9999
16 -9999
17 -9999
```

Preparation of settings file

All in- and output files need to be defined in the settings file. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element. Just make sure that the map with the polder locations is in the "maps" directory, and all tables in the 'tables" directory. If this is the case, you only have to specify the initial reservoir water level in the polders. *PolderInitialLevelValue* is defined in the 'lfuser' element of the settings file, and it can be either a map or a value. The value of the weir constant is also defined here, although you should not change its default value. So we add this to the 'lfuser' element (if it is not there already):

```
<group>
<comment>
************************************************************
POLDER OPTION
************************************************************
</comment>
<textvar name="mu" value="0.49">
<comment>
Weir constant [-] (Do not change!)
</comment>
</textvar>
<textvar name="PolderInitialLevelValue" value="0">
<comment>
Initial water level in polder [m]
</comment>
</textvar>
</group>
```

To switch on the polder routine, add the following line to the 'lfoptions' element:

```
<setoption name="simulatePolders" choice="1" />
```

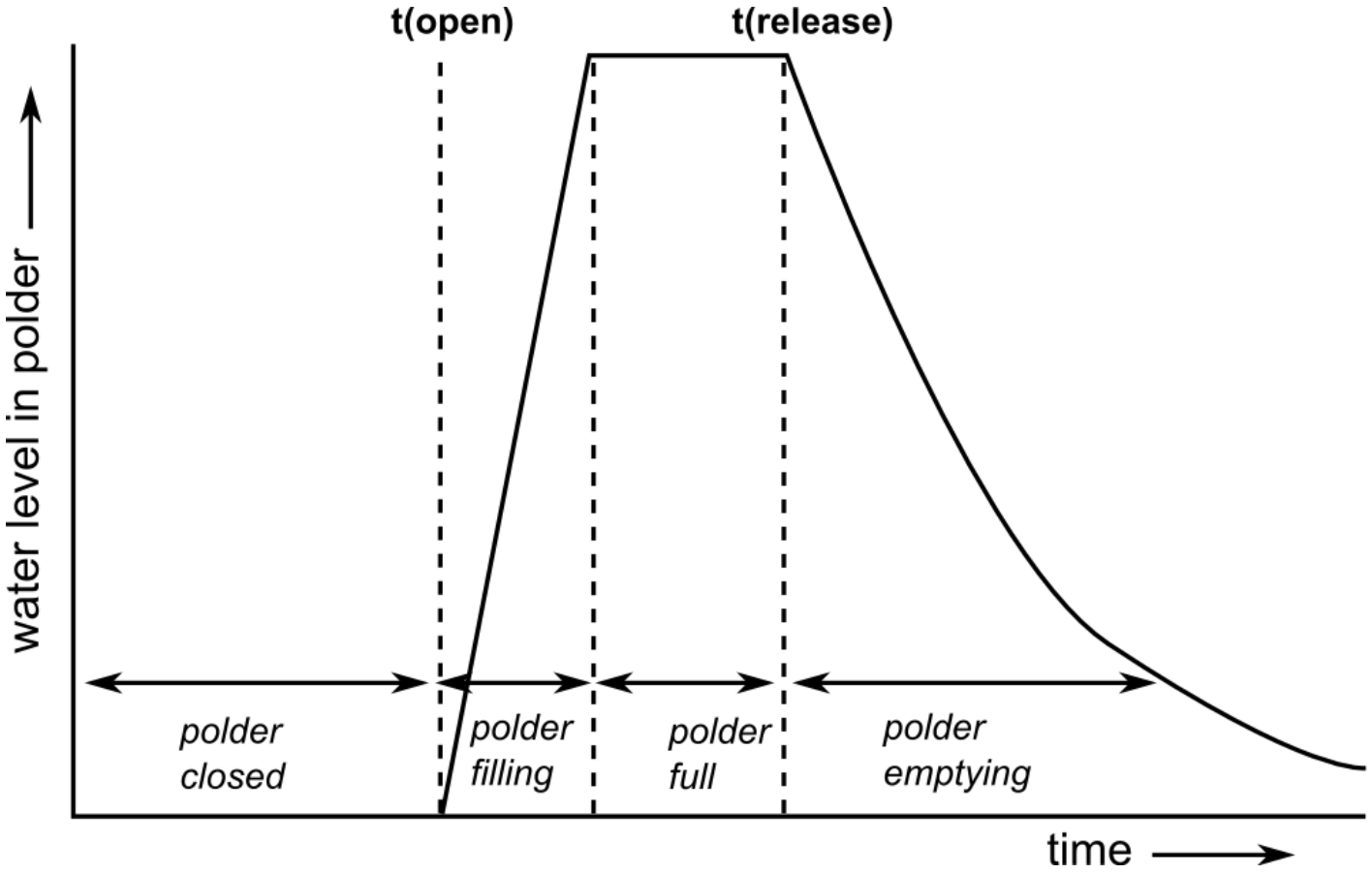

t(open)
t(release)
water level in polder
polder closed
polder filling
polder full
polder emptying
time

Figure 33:

Now you are ready to run the model. If you want to compare the model results both with and without the inclusion of polders, you can switch off the simulation of polders either by:

1. Removing the 'simulatePolders' statement from the 'lfoptions element, or
2. changing it into <setoption name="simulatePolders" choice="0" />

Both have exactly the same effect. You don't need to change anything in either 'lfuser' or 'lfbinding'; all file definitions here are simply ignored during the execution of the model.

Polder output files

The polder routine produces 2 additional time series and one map (or stack of maps, depending on the value of LISFLOOD variable *ReportSteps*), as listed in the following table:

Table: *Output of polder routine.*

Maps / Time series	Default name	Description	Units
PolderLevelState	hpolxxxx.xxx	water level in polder at last time step	m
PolderLevelTS	hPolder.tss	water level in polder (at polder locations)	m
PolderFluxTS	qPolder.tss	Flux into and out of polder (positive for flow from channel to polder, negative for flow from polder to channel)	$\frac{m^3}{s}$

Note that you can use the map with the polder level at the last time step to define the initial conditions of a succeeding simulation, e.g.:

```
<textvar name="PolderInitialLevelValue" value="/mycatchment/hpol0000.730">
```

Limitations

For the moment, polders can be simulated on channel pixels where dynamic wave routing is used. For channels where the kinematic wave is used, the routine will not work and may lead to numerical instabilities or even model crashes. This limitation may be resolved in future model versions. Has that been fixed in the meantime?

Simulation of reservoirs

Introduction

This page describes the LISFLOOD reservoirs routine, and how it is used. The simulation of reservoirs is *optional*, and it can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="simulateReservoirs" choice="1" />
```

Reservoirs can be simulated on channel pixels where kinematic wave routing is used. The routine does *not* work for channel stretches where the dynamic wave is used!

Description of the reservoir routine

Reservoirs are simulated as points in the channel network. The inflow into each reservoir equals the channel flow upstream of the reservoir. The outflow behaviour is described by a number of parameters. First, each reservoir has a total storage capacity S $[m^3]$. The relative filling of a reservoir, F, is a fraction between 0 and 1. There are three 'special' filling levels. - L_c: 'conservative storage limit'. This is the lower limit as reservoirs are never completely empty. - L_f: 'flood storage limit'. This is the upper limit as reservoirs are never filled completly for safety reasons - L_n: is the available capacity of a reservoir between L_f and L_c.

Three additional parameters define the way the outflow of a reservoir is regulated. - 'minimum outflow' (O_{min}, $[\frac{m^3}{s}]$) which is maintained for e.g. ecological reasons; - 'non-damaging outflow' (O_{nd}, $[\frac{m^3}{s}]$) is the maximum possible outflow that will not cause problems downstream; and - 'normal outflow' (O_{norm}, $[\frac{m^3}{s}]$) is the one valid when the reservoir is within its 'normal storage' filling level.

Depending on the relative filling of the reservoir, outflow ($O_{res}, [\frac{m^3}{s}]$) is calculated as:

If

$$F \leq 2 \cdot L_c$$

, then:

$$O_{res} = min(O_{min}, \frac{1}{\Delta t} \cdot F \cdot S)$$

If

$$L_n \geq F > 2L_c$$

, then:

$$O_{res} = O_{min} + (O_{norm} - O_{min}) \cdot \frac{(F - 2L_c)}{(L_n - 2L_c)}$$

If

$$L_f \geq F > L_n$$

, then:

$$O_{res} = O_{norm} + \frac{(F - L_n)}{(L_f - L_n)} \cdot \max((I_{res} - O_{norm}), (O_{nd} - O_{norm}))$$

If

$$F > L_f$$

, then:

$$O_{res} = \max(\frac{(F - L_f)}{\Delta t} \cdot S, O_{nd})$$

with: S: Reservoir storage capacity $[m^3]$ F: Reservoir fill (fraction, 1 at total storage capacity)

–

L_c: Conservative storage limit

–

L_n: Normal storage limit

–

L_f: Flood storage limit

–

O_{min}: Minimum outflow $[\frac{m^3}{s}]$ O_{norm}: Normal outflow $[\frac{m^3}{s}]$ O_{nd}: Non-damaging outflow $[\frac{m^3}{s}]$ I_{res}: Reservoir inflow $[\frac{m^3}{s}]$

In order to prevent numerical problems, the reservoir outflow is calculated using a user-defined time interval (or Δt, if it is smaller than this value).

Preparation of input data

For the simulation of reservoirs a number of additional input files are necessary. First, the locations of the reservoirs are defined on a (nominal) map called '*res.map*'. It is important that all reservoirs are located on a channel pixel (you can verify this by displaying the reservoirs map on top of the channel map). Also, since each reservoir receives its inflow from its upstream neighbouring channel pixel, you may want to check if each reservoir has any upstream channel pixels at all (if not, the reservoir will gradually empty during a model run!). The management of the reservoirs is described by 7 tables. The following table lists all required input:

Table: *Input requirements reservoir routine.*

Maps	Default name	Description	Units	Remarks
ReservoirSites	res.map	reservoir locations	-	nominal
TabTotStorage	rtstor.txt	reservoir storage capacity	$[m^3]$	
TabConservativeStorageLimit	rclim.txt	conservative storage limit	-	fraction of storage
TabNormalStorageLimit	rnlim.txt	normal storage limit	-	capacity
TabFloodStorageLimit	rflim.txt	flood storage limit	-	
TabMinOutflowQ	rminq.txt	minimum outflow	$[m^3]$	
TabNormalOutflowQ	rnormq.txt	normal outflow	$[m^3]$	
TabNonDamagingOutflowQ	rndq.txt	non-damaging outflow	$[m^3]$	

When you create the map with the reservoir sites, pay special attention to the following: if a reservoir is on the most downstream cell (i.e. the outflow point, see Figure below), the reservoir routine may produce erroneous output. In particular, the mass balance errors cannot be calculated correctly in that case. The same applies if you simulate only a sub-catchment of a larger map (by selecting the subcatchment in the mask map). This situation can usually be avoided by extending the mask map by one cell in downstream direction.

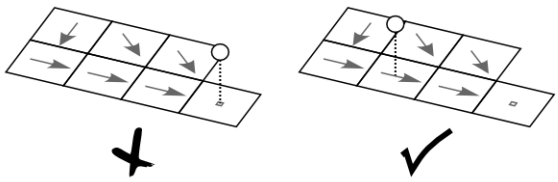


Figure 34:

Figure: *Placement of the reservoirs: reservoirs on the outflow point (left) result in erroneous behavior of the reservoir routine.*

Preparation of settings file

All in- and output files need to be defined in the settings file. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element. Just make sure that the map with the reservoir locations is in the "maps" directory, and all tables in the 'tables" directory. If this is the case, you only have to specify the time-step used for the reservoir calculations and the initial reservoir filling level (expressed as a fraction of the storage capacity). Both are defined in the 'lfuser' element of the settings file. For the reservoir time step (DtSecReservoirs) it is recommended to start with a value of 14400 (4 hours), and try smaller values if the simulated reservoir outflow shows large oscillations. ReservoirInitialFillValue can be either a map or a value (between 0and 1). So we add this to the 'lfuser' element (if it is not there already):

```
<group>
<comment>
```

```
**********************************************************
RESERVOIR OPTION
**********************************************************
</comment>
<textvar name="DtSecReservoirs" value="14400">
<comment>
Sub time step used for reservoir simulation [s]. Within the model,
the smallest out of DtSecReservoirs and DtSec is used.
</comment>
</textvar>
<textvar name="ReservoirInitialFillValue" value="-9999">
<comment>
Initial reservoir fill fraction
-9999 sets initial fill to normal storage limit
if you're not using the reservoir option, enter some bogus value
</comment>
</textvar>
</group>
```

The value -9999 tells the model to set the initial reservoir fill to the normal storage limit. Note that this value is completely arbitrary. However, if no other information is available this may be a reasonable starting point.

Finally, you have to tell LISFLOOD that you want to simulate reservoirs! To do this, add the following statement to the 'lfoptions' element:

```
<setoption name="simulateReservoirs" choice="1" />
```

Now you are ready to run the model. If you want to compare the model results both with and without the inclusion of reservoirs, you can switch off the simulation of reservoirs either by:

1. Removing the 'simulateReservoirs' statement from the 'lfoptions element, or
2. changing it into <setoption name="simulateReservoirs" choice="0" />

Both have exactly the same effect. You don't need to change anything in either 'lfuser' or 'lfbinding'; all file definitions here are simply ignored during the execution of the model.

Reservoir output files

The reservoir routine produces 3 additional time series and one map, as listed in the following table:

Table: *Output of reservoir routine.*

Maps / Time series	Default name	Description	Units	Remarks
ReservoirFillState	rsfilxxx.xxx	reservoir fill at last time step[^10]	-	
ReservoirInflowTS	qresin.tss	inflow into reservoirs	$\frac{m^3}{s}$	
ReservoirOutflowTS	qresout.tss	outflow out of reservoirs	$\frac{m^3}{s}$	
ReservoirFillTS	resfill.tss	reservoir fill	-	

Note that you can use the map with the reservoir fill at the last time step to define the initial conditions of a succeeding simulation, e.g.:

```
<textvar name="ReservoirInitialFillValue" value="/mycatchment/rsfil000.730">
```

Simulation and reporting of soil moisture as pF values

Introduction

LISFLOOD offers the possibility to calculate pF values from the moisture content of both soil layers. The calculation of pF values is *optional*, and it can be activated by adding the following line to the 'lfoptions' element in the LISFLOOD settings file (lfoptions):

```
<setoption name="simulatePF" choice="1" />
```

Using this option does *not* influence the actual model results in any way, and it is included only to allow the model user to report pF time series or maps. The actual *reporting* of the computed pF values (as time series or maps) can be activated using separate options (which are discussed further down).

Calculation of pF

A soil's pF is calculated as the logarithm of the capillary suction head, h:

$$pF = \log_{10}(h)$$

with h in

$$cm$$

(positive upwards). Values of pF are typically within the range 1.0 (very wet) to 5.0 (very dry). The relationship between soil moisture status and capillary suction head is described by the Van Genuchten equation (here again re-written in terms of mm water slice, instead of volume fractions):

$$h = \frac{1}{\alpha}[(\frac{w_s - w_r}{w - w_r})^{\frac{1}{m}} - 1]^{\frac{1}{n}}$$

where h is the suction head $[cm]$, and w, w_r and w_s are the actual, residual and maximum amounts of moisture in the soil respectively (all in mm). Parameter is related to soil texture. Parameters m and n are calculated from the pore-size index, (which is related to soil texture as well):

$$m = \frac{\lambda}{\lambda + 1}$$

$$n = \lambda + 1$$

If the soil contains no moisture at all (w=0), h is set to a fixed (arbitrary) value of 110^7 cm.

Reporting of pF

pF can be reported as time series (at the locations defined on the "sites" map or as average values upstream each gauge location), or as maps. To generate time series at the "sites", add the following line to the 'lfoptions' element of your settings file:

```
<setoption name=\"repPFTs\" choice=\"1\" />
```

For maps, use the following lines instead (for the upper and lower soil layer, respectively):

```
<setoption name=\"repPF1Maps\" choice=\"1\" />

<setoption name=\"repPF2Maps\" choice=\"1\" />
```

In either case, the reporting options should be used *in addition* to the 'simulatePF' option. If you do not include the 'simulatePF' option, there will be nothing to report and LISFLOOD will exit with an error message.

Preparation of settings file

The naming of the reported time series and maps is defined in the settings file. The two Tables at the end of this page list the settings variables default output names. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element.

Time series:

```
<comment>
PF TIMESERIES, VALUES AT SITES
</comment>
<textvar name="PF1TS" value="$(PathOut)/pFTop.tss">
<comment>
Reported pF upper soil layer [-]
</comment>
</textvar>
<textvar name="PF2TS" value="$(PathOut)/pFSub.tss">
<comment>
Reported pF lower soil layer [-]
</comment>
</textvar>
<comment>
PF TIMESERIES, AVERAGE VALUES UPSTREAM OF GAUGES
</comment>
<textvar name="PF1AvUpsTS" value="$(PathOut)/pFTopUps.tss">
<comment>
Reported pF upper soil layer [-]
</comment>
</textvar>
<textvar name="PF2AvUpsTS" value="$(PathOut)/pFSubUps.tss">
<comment>
Reported pF lower soil layer [-]
</comment>
</textvar>
```

Map stacks:

```
<comment>
PF MAPS
</comment>
<textvar name="PF1Maps" value="$(PathOut)/pftop">
<comment>
Reported pF upper soil layer [-]
</comment>
</textvar>
<textvar name="PF2Maps" value="$(PathOut)/pfsub">
<comment>
Reported pF lower soil layer [-]
</comment>
</textvar>
```

Table: *pF map output*

Description	Option name	Settings variable	Default prefix
pF upper layer	repPF1Maps	PF1Maps	pftop
pF lower layer	repPF2Maps	PF2Maps	pfsub

Table: *pF timeseries output*

Description	Settings variable	Default name
pF at sites (option repPFSites)		
pF upper layer	PF1TS	pFTop.tss
pF lower layer	PF2TS	pFSub.tss
pF, average upstream of gauges (option repPFUpsGauges)		
pF upper layer	PF1AvUpsTS	pFTopUps.tss
pF lower layer	PF2AvUpsTS	pFSubUps.tss

Transmission loss option

Introduction

This page describes the LISFLOOD transmission loss routine, and how it is used.

The term 'transmission loss' originate from electronic or communication science and stands for: "The loss of power or voltage of a transmitted wave or current in passing along a transmission line or path or through a circuit device". In river systems, particularly in semi-arid and arid region a similar effect can be observed: The loss of water along river channel especially during low and average flow periods. Main reasons for this loss might be:

- Evaporation of water inside the channel reach
- Use of water for domestic, industrial or agricultural use
- Infiltration to lower groundwater zones

A simplified approach to model this effect has been chosen from Rao and Maurer (1996), without the need of additional data and with only three parameters, making calibration relatively simple.

Transmission loss is *optional*, and can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="TransLoss" choice="1" />
```

Description of the transmission loss approach

The approach by Rao and Maurer 1996 builds a one-parameter relationship between the seepage of a channel with the depth of flow. A power relationship is then utilized for the stage-discharge relationship, which is coupled with the seepage relationship.

Someone please add the equation… I can't find it. Thanks

$$Outflow = xxx$$

with: $Outflow$: discharge at the outflow $Inflow$: discharge at the Inflow (upstream) $TransPower$: Parameter given by the rating curve $TransSub$: Parameter which is to calibrate

As a main difference to the Rao and Maurer 1996, the $TransPower$ parameter is not calculated by using a rating curve but is estimated (calibrated) as the parameter $TransSub$. Transmission loss takes place where the channel gets bigger with more influence of river-aquifer interaction and also with more river-floodplain interaction. Therefore a minimum upstream area is defined where transmission loss starts to get important.

Using transmission loss

No additional maps or tables are needed. Using the transmission loss option involves two steps:

1) In the 'lfuser' element (replace the file paths/names by the ones you want to use):

```
<group>
<comment>
************************************************************
TRANSMISSION LOSS PARAMETERS
************************************************************
Suggested parameterisation strategy:
Use TransSub as calibration constant leave all other parameters at\
default values
</comment>
<textvar name="TransSub" value="0.3">
<comment>
Transmission loss function parameter
Standard: 0.3 Range: 0.1 - 0.6
</comment>
</textvar>
<textvar name="TransPower1" value="2.0">
<comment>
Transmission loss function parameter
Standard: 2.0 Range 1.3 -- 2.0
</comment>
</textvar>
<textvar name="TransArea" value="5.0E+10">
<comment>
downstream area taking into account for transmission loss
Standard: 5.0E+10 Range: 1.0E+10 -- 1.0E+11
</comment>
</textvar>
</group>
```

TransSub is the linear transmission loss parameter. Standard is set to 0.3 and the range should be between 0.1 – 0.6 (higher values lead to more loss)

TransPower is the power transmission loss parameter. Standard is set to 2.0 and the range should be between 1.3 and 2.0 (higher values lead to more loss)

TransArea is the downstream area which is taken into account for transmission loss. Standard is $5.0E+10km^2$ and range should be $1.0E+10km^2$ to $1.0E+11km^2$ (higher values lead to less loss as less area is taken into account)

2) Activate the transmission loss option

Add the following line to the 'lfoptions' element:

```
<setoption name="TransLoss" choice="1" />
```

Now you are ready to run the model with the transmission loss option

Transmission loss output file

The transmission option can produce an additional time series as listed in the following table:

Table: *Output of transmission loss routine – Average upstream of gauges.*

Time series	Default name	Description	Units
TransLossAvUps	TransLossAvUps.tss	Transmission loss in the channel	*mm*

Simulation and reporting of water levels

Introduction

Within LISFLOOD it is possible to simulate and report water levels in the river channel. This page describes the LISFLOOD water levels option, and how it is used. The simulation of water levels is *optional*, and it can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="simulateWaterLevels" choice="1" />
```

If the option is switched on, water levels are calculated for river channel pixels where either kinematic or dynamic wave routing is used. Using this option does *not* influence the actual model results in any way, and it is included only to allow the model user to report water levels. The actual *reporting* of the simulated water levels (as time series or maps) can be activated using two separate options (see below in section 'Reporting of water levels'):

Calculation of water levels

For channel stretches that are simulated using the dynamic wave, the water level in the channel is simply the difference between the channel head and the channel bottom level. For kinematic wave stretches, only approximate water levels can be estimated from the cross-sectional (wetted) channel area, A_{ch} for each time step. Since the channel cross-section is described as a trapezoid, water levels follow directly from A_{ch} , channel bottom width, side slope and bankfull level. If A_{ch} exceeds the bankfull cross-sectional area (A_{bf}), the surplus is distributed evenly over the (rectangular) floodplain, and the depth of water on the floodplain is added to the (bankfull) channel depth. The Figure below further illustrates the cross-section geometry. All water levels are relative to channel bottom level (z_{bot} in the Figure).

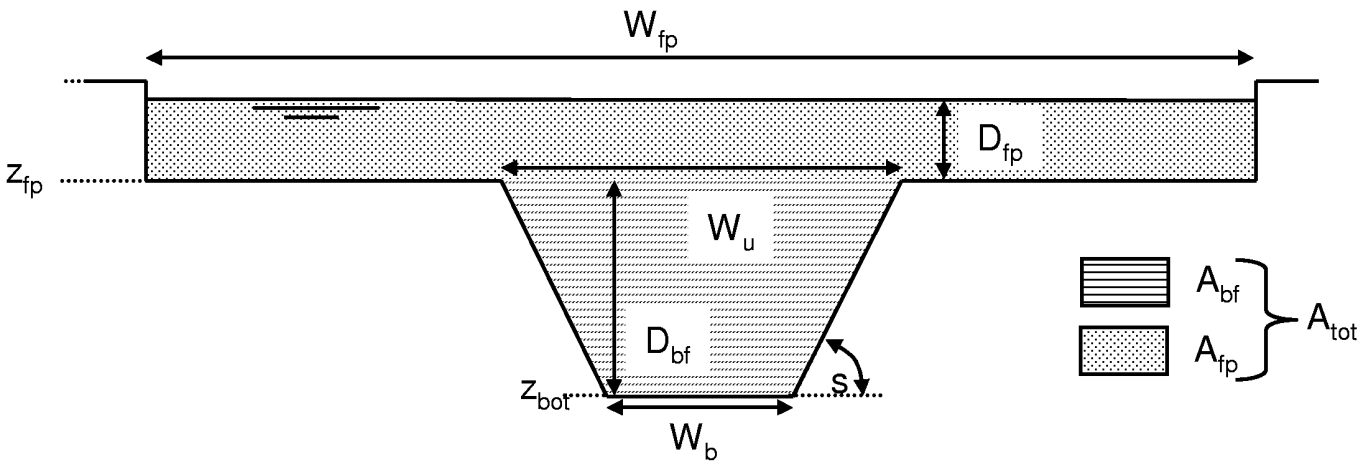


Figure 35:

Figure: *Geometry of channel cross-section in kinematic wave routing. With W_b: channel bottom width; W_u: channel upper width; z_{bot}: channel bottom level; z_{fp}: floodplain bottom level; s: channel side slope; W_{fp}: floodplain width; A_{bf}: channel cross-sectional area at bankfull; A_{fp}: floodplain cross-sectional area; D_{bf}: bankfull channel depth, D_{fp}: depth of water on the floodplain.*

In order to calculate water levels, LISFLOOD needs a map with the width of the floodplain in

$$m$$

, which is defined by 'lfbinding' variable *FloodPlainWidth* (the default name of this map is *chanfpln.map*).

Reporting of water levels

Water levels can be reported as time series (at the gauge locations that are also used for reporting discharge), or as maps.

To generate a time series, add the following line to the 'lfoptions' element of your settings file:

```
<setoption name="repWaterLevelTs" choice="1" />
```

For maps, use the following line instead:

```
<setoption name="repWaterLevelMaps" choice="1" />
```

In either case, the reporting options should be used *in addition* to the 'simulateWaterLevels' option. If you do not include the 'simulateWaterLevels' option, there will be nothing to report and LISFLOOD will exit with an error message.

Preparation of settings file

The naming of the reported water level time series and maps is defined in the settings file. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element.

Time series:

```
<textvar name="WaterLevelTS" value="$(PathOut)/waterLevel.tss">
<comment>
Reported water level [m]
</comment>
</textvar>
```

Map stack:

```
<textvar name="WaterLevelMaps" value="$(PathOut)/wl">
<comment>
Reported water level [m]
</comment>
</textvar>
```

Including water use

Introduction

This page describes the LISFLOOD water use routine, and how it is used.

The water use routine can be used to assess the effect of water withdrawal from different sectors to the water balance. Sectors can be public, industrial, agricultural or energy water withdrawal. As input LISFLOOD needs a stack of maps for one representative year of total water withdrawal demand. LISFLOOD calculates if the available water can fulfill the sectorial water demand and if yes, withdrawals the amount of water from the river system.

Water use is *optional*, and can be activated by adding the following line to the 'lfoptions' element:

```
<setoption name="wateruse" choice="1" />
```

Calculation of water use

The water is withdrawn only from discharge in the river network but not from soil, groundwater or directly from precipitation.

- For each single day a total demand of withdrawal water is loaded from a sparse stack of maps
- Water use is taken from the discharge in the river network. First the water use is taken from the same grid cell (see figure below – pixel No. 1)
- If the amount of water withdrawal is larger than the water available in this grid cell water is taken from downstream moving along the local drain direction. This is done by implementing a loop substracting the remaining water from the next downstream cell till all the water for water use is taken or a predefined number of iteration is reached (see figure below – pixel No. 2 to 5)

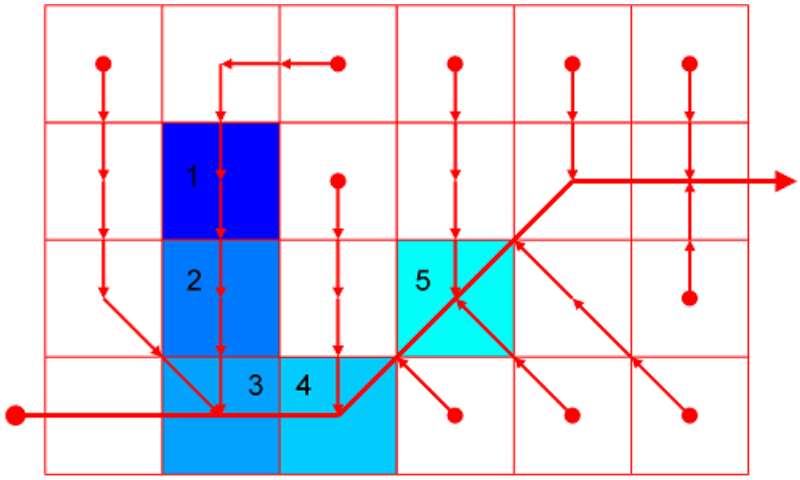


Figure 36:

Figure: *Water withdrawal assessing demand and availability along the flow path.*

In the LISFLOOD settings file you can define:

- the percentage of water that must remain in a grid cell and is not withdrawn by water use (WUsePercRemain)
- the maximum number of loops (= distance to the water demand cell). For example in figure above: maxNoWateruse = 5

Preparation of input data

The following Table gives an overview about the maps and table needed for the water use option.

Table: *Input requirements water use.*

Maps	Default name	Description	Units
Yearly stack of water use maps	wuse0000.xxx	Total water withdrawal	$\frac{m^3}{s}$
Table WUseofDay	WUseofDays.txt	Assignment of day of the year to map stack of water use	-

A sparse map stack of one year of total water withdrawal [$\frac{m^3}{s}$] with a map every 10 days or a month is needed. Because it is assumed that water use follows a yearly circle, this map stack is used again and again for the following years. For example:

```
t   map name
1   wuse0000.001
2   wuse0000.032
3   wuse0000.060
4   wuse0000.091
5   wuse0000.121
6   wuse0000.152
7   wuse0000.182
8   wuse0000.213
9   wuse0000.244
10  wuse0000.274
11  wuse0000.305
12  wuse0000.335
```

To assign each day of simulation the right map a lookup table (WUseOfDay.txt) is necessary:

```
<,0.5> 335
[0.5,32> 1
[32,60> 32
```

Preparation of settings file

All in- and output files need to be defined in the settings file. If you are using a default LISFLOOD settings template, all file definitions are already defined in the 'lfbinding' element.

```
<textvar name="PathWaterUse" value="./wateruse">
<comment>
Water use maps path
</comment>
</textvar>
<textvar name="PrefixWaterUse" value="wuse">
<comment>
prefix water use maps
</comment>
</textvar>
<group>
<comment>
***************************************************************
INPUT WATER USE MAPS AND PARAMETERS
***************************************************************
</comment>
<textvar name="WaterUseMaps" value="$(PathOut)/wuse">
<comment>
Reported water use [cu m/s], depending on the availability of
discharge
</comment>
</textvar>
<textvar name="WaterUseTS" value="$(PathOut)/wateruseUps.tss">
<comment>
Time series of upstream water use at gauging stations
</comment>
</textvar>
<textvar name="StepsWaterUseTS"
value="$(PathOut)/stepsWaterUse.tss">
<comment>
Number of loops needed for water use
routine
</comment>
</textvar>
<textvar name="maxNoWateruse" value="5">
<comment>
maximum number of loops for calculating the use of water
= distance water is taken for water consuption
</comment>
</textvar>
<textvar name="WUsePercRemain" value="0.2">
<comment>
percentage of water which remains in the channel
e.g. 0.2 -> 20 percent of discharge is not taken out
</comment>
</textvar>
</group>
```

PathWaterUse is the path to the map stack of water use

PrefixWaterUse is the prefix of the water use maps

WaterUseMaps is the path and prefix of the reported water use $[\frac{m^3}{s}]$ as a result of demand and availability

WaterUseTS are time series of upstream water use $[\frac{m^3}{s}]$ at gauging stations

StepsWaterUseTS is the number of loops needed for water use [-]

maxNoWateruse is maximum number of loops for calculating the use of water (distance water is taken for water consumption)

WUsePercRemain is the percentage of water which remains in the channel (e.g. 0.2 -> 20 percent of discharge is not taken out)

Finally, you have to tell LISFLOOD that you want to simulate water use. To do this, add the following statement to the 'lfoptions' element:

```
<setoption name="wateruse" choice="1" />
```

Water use output files

The water use routine can produce 2 additional time series and one map (or stack), as listed in the following Table:

Table: *Output of water use routine.*

Maps	Default name	Option	Units
WaterUseMaps	wusexxxx.xxx	repwateruseMaps	$\frac{m^3}{s}$
Time series			
Number of loops	stepsWaterUse.tss		-
WaterUseTS	wateruseUps.tss	repwateruseTS	$\frac{m^3}{s}$

LISFLOOD input files

All input that LISFLOOD requires are either in map or table format. Before showing a listing of all LISFLOOD input files, first some important remarks on the meteorological input data LISFLOOD requires.

Treatment of meteorological input variables

The meteorological conditions provide the driving forces behind the water balance. LISFLOOD uses the following meteorological input variables:

Code	**Description**	**Unit**
P	Precipitation	$[\frac{mm}{day}]$
$ET0$	Potential (reference) evapotranspiration rate	$[\frac{mm}{day}]$
$EW0$	Potential evaporation rate from open water surface	$[\frac{mm}{day}]$
$ES0$	Potential evaporation rate from bare soil surface	$[\frac{mm}{day}]$
T_{avg}	Average *daily* temperature	$^{\circ}C$

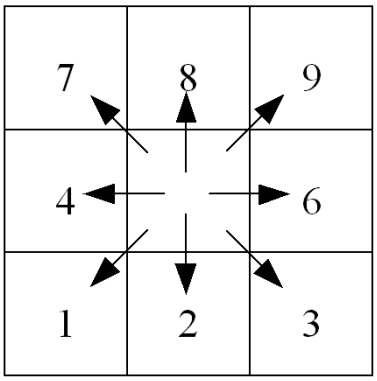
> **Note** that the model needs *daily* average temperature values, even if the model is run on a smaller time interval (e.g. hourly). This is because the routines for snowmelt and soil freezing are use empirical relations which are based on daily temperature data **Cinzia is that still correct??**. Just as an example, feeding hourly temperature data into the snowmelt routine can result in a gross overestimation of snowmelt. This is because even on a day on which the average temperature is below T_m (no snowmelt), the instantaneous (or hourly) temperature may be higher for a part of the day, leading to unrealistically high simulated snowmelt rates.

Both precipitation and evaporation are internally converted from *intensities* $[\frac{mm}{day}]$ to *quantities per time step* $[mm]$ by multiplying them with the time step, Δt (in *days*). For the sake of consistency, all in- and outgoing fluxes will also be described as *quantities per time step* $[mm]$ in the following, unless stated otherwise. $ET0$, $EW0$ and $ES0$ can

be calculated using standard meteorological observations. To this end a dedicated pre-processing application has been developed (LISVAP), which is documented in a separate manual. Insert link to LISVAP manual once it is produced

LISFLOOD input maps

Table: *LISFLOOD input maps.*

Map	Default name	Units, range	Description
GENERAL			
MaskMap	area.map	Unit: - Range: 0 or 1	Boolean map that defines model boundaries
TOPOGRAPHY			
Ldd	ldd.map	U.: flow directions R.: 1 map 9	local drain direction map (with value 1-9); this file contains flow directions from each cell to its steepest downslope neighbour. Ldd directions are coded according to the following diagram:  This resembles the numeric key pad of your PC's keyboard, except for the value 5, which defines a cell without local drain direction (pit). The pit cell at the end of the path is the outlet point of a catchment.
Grad	gradient.map	U.: $\frac{m}{m}$ R.: map > 0 !!!	Slope gradient
Elevation Stdev	elvstd.map	U.: m R.: map 0	Standard deviation of elevation
LAND USE – fraction maps			
Fraction of water	fracwater.map	U.: [-] R.: 0 map 1	Fraction of inland water for each cell. Values range from 0 (no water at all) to 1 (pixel is 100% water)
Fraction of sealed surface	fracsealed.map	U.: [-] R.: 0 map 1	Fraction of impermeable surface for each cell. Values range from 0 (100% permeable surface – no urban at all) to 1 (100% impermeable surface).
Fraction of forest	fracforest.map	U.:[-] R.: 0 map 1	Forest fraction for each cell. Values range from 0 (no forest at all) to 1 (pixel is 100% forest)

Map	Default name	Units, range	Description
Fraction of other land cover	fracother.map	U.: [-] R.: 0 map 1	Other (agricultural areas, non-forested natural area, pervious surface of urban areas) fraction for each cell.
LAND COVER depending maps			
Crop coef. for forest	cropcoef_forest.map	U.: [-] R.: 0.8 map 1.2	Crop coefficient for forest
Crop coef. for other	cropcoef_other.map	U.: [-] R.: 0.8 map 1.2	Crop coefficient for other
Crop group number for forest	crgrnum_forest.map	U.: [-] R.: 1 map 5	Crop group number for forest
Crop group number for forest	crgrnum_other.map	U.: [-] R.: 1 map 5	Crop group number for other
Manning for forest	mannings_forest.map	U.: [-] R.: 0.2 map 0.4	Manning's roughness for forest
Manning for other	mannings_other.map	U.: [-] R.: 0.01 map 0.3	Manning's roughness for other
Soil depth for forest for layer1	soildep1_forest.map	U.: mm R.: map 50	Forest soil depth for soil layer 1 (rooting depth)
Soil depth for other for layer1	soildep1_other.map	U.: mm R.: map 50	Other soil depth for soil layer 1 (rooting depth)
Soil depth for forest for layer2	Soildep2_forest.map	U.: mm R.: map 50	Forest soil depth for soil layer 2
Soil depth for other for layer2	Soildep2_other.map	U.: mm R.: map 50	Other soil soil depth for soil layer 2
SOIL HYDRAULIC PROPERTIES (depending on soil texture)			
ThetaSat1 for forest	thetas1_forest.map	U.: [-] R.: $0 < \text{map} < 1$	Saturated volumetric soil moisture content layer 1
ThetaSat1 for other	thetas1_other.map	U.: [-] R.: $0 < \text{map} < 1$	Saturated volumetric soil moisture content layer 1
ThetaSat2 for forest and other	thetas2.map	U.: [-] R.: $0 < \text{map} < 1$	Saturated volumetric soil moisture content layer 2
ThetaRes1 for forest	thetar1_forest.map	U.: [-] R.: $0 < \text{map} < 1$	Residual volumetric soil moisture content layer 1
ThetaRes1 for other	thetar1_other.map	U.: [-] R.: $0 < \text{map} < 1$	Residual volumetric soil moisture content layer 1
ThetaRes2 for forest and other	thetar2.map	U.: [-] R.: $0 < \text{map} < 1$	Residual volumetric soil moisture content layer 2
Lambda1 for forest	lambda1_forest.map	U.: [-] R.: $0 < \text{map} < 1$	Pore size index ( ) layer 1
Lambda1 for other	lambda1_other.map	U.: [-] R.: $0 < \text{map} < 1$	Pore size index ( ) layer 1
Lambda2 for forest and other	lambda2.map	U.: [-] R.: $0 < \text{map} < 1$	Pore size index ( ) layer 2
GenuAlpha1 for forest	alpha1_forest.map	U.: [-] R.: $0 < \text{map} < 1$	Van Genuchten parameter layer 1
GenuAlpha1 for other	alpha1_other.map	U.: [-] R.: $0 < \text{map} < 1$	Van Genuchten parameter layer 1
GenuAlpha2 for forest and other	alpha2.map	U.: [-] R.: $0 < \text{map} < 1$	Van Genuchten parameter layer 2
Sat1 for forest	ksat1_forest.map	U.: $\frac{cm}{day}$ R.: 1 map 100	Saturated conductivity layer 1
Sat1 for other	ksat1_other.map	U.: $\frac{cm}{day}$ R.: 1 map 100	Saturated conductivity layer 1
Sat2 for forest and other	ksat2.map	U.: $\frac{cm}{day}$ R.: 1 map 100	Saturated conductivity layer 2
CHANNEL GEOMETRY			
Channels	chan.map	U.: [-] R.: 0 or 1	Map with Boolean 1 for all channel pixels, and Boolean 0 for all other pixels on MaskMap
ChanGrad	changrad.map	U.: $\frac{m}{m}$ R.: map > 0 !!!	Channel gradient
ChanMan	chanman.map	U.: [-] R.: map > 0	Manning's roughness coefficient for channels
ChanLength	chanleng.map	U.: m R.: map > 0	Channel length (can exceed grid size, to account for meandering rivers)

Map	Default name	Units, range	Description
ChanBottomWidth	chanbw.map	U.: m R.: map > 0	Channel bottom width
ChanSdXdY	chans.map	U.: $\frac{m}{m}$ R.: map $\quad 0$	Channel side slope Important: defined as horizontal divided by vertical distance (dx/dy); this may be confusing because slope is usually defined the other way round (i.e. dy/dx)!
ChanDepthThreshold	chanbnkf.map	U.: m R.: map > 0	Bankfull channel depth
METEOROLOGICAL VARIABLES			
PrecipitationMaps	pr	U.: $\frac{mm}{day}$ R.: map $\quad 0$	Precipitation rate
TavgMaps	ta	U.: $\check{r}C$ R.:-50 map $\quad +50$	Average daily temperature
E0Maps	e	U.: $\frac{mm}{day}$ R.: map $\quad 0$	Daily potential evaporation rate, free water surface
ES0Maps	es	U.: $\frac{mm}{day}$ R.: map $\quad 0$	Daily potential evaporation rate, bare soil
ET0Maps	et	U.: $\frac{mm}{day}$ R.: map $\quad 0$	Daily potential evapotranspiration rate, reference crop
DEVELOPMENT OF VEGETATION OVER TIME			
LAIMaps for forest	lai_forest	U.: $\frac{m^2}{m^2}$ R.: map $\quad 0$	Pixel-average Leaf Area Index for forest
LAIMaps for other	lai_other	U.: $\frac{m^2}{m^2}$ R.: map $\quad 0$	Pixel-average Leaf Area Index for other
DEFINITION OF INPUT/OUTPUT TIMESERIES			
Gauges	outlets.map	U.: [-] R.: For each station an individual number	Nominal map with locations at which discharge timeseries are reported (usually correspond to gauging stations)
Sites	sites.map	U.: [-] R.: For each station an individual number	Nominal map with locations (individual pixels or areas) at which timeseries of intermediate state and rate variables are reported (soil moisture, infiltration, snow, etcetera)

Table: *Optional maps that define grid size.*

Map	Default name	Units, range	Description
PixelLengthUser	pixleng.map	U.: m R.: map > 0	Map with pixel length
PixelAreaUser	pixarea.map	U.: m R.: map > 0	Map with pixel area

Tables

Is this section up to date??

In the previous version of LISFLOOD a number of model parameters are read through tables that are linked to the classes on the land use and soil (texture) maps. Those tables are replaced by maps (e.g. soil hydraulic property maps) in order to include the sub-grid variability of each parameter. Therefore only one default table is used in the standard LISFLOOD

setting. The following table gives an overview:

Table: *LISFLOOD input tables.*

Table	Default name	Description
LAND USE		
Day of the year -> LAI	LaiOfDay.txt	Lookup table: Day of the year -> LAI map

Output generated by LISFLOOD

Default LISFLOOD output

LISFLOOD can generate a wide variety of output. Output is generated as either maps or time series (PCRaster format, which can be visualised with PCRaster's 'aguila' application; or as NetCDF). Reporting of output files can be switched on and off using options in the LISFLOOD settings file. Also, a number of output files are specific to other optional modules, such as the simulation of reservoirs. The following table lists all the output time series that are reported by default (note that the file names can always be changed by the user, although this is not recommended):

Table: *LISFLOOD default output time series.*

Description	Units	File name
RATE VARIABLES AT GAUGES		
[1,2] channel discharge	$\frac{m^3}{s}$	dis.tss
NUMERICAL CHECKS		
[2] cumulative mass balance error	m^3	mbError.tss
[2] cumulative mass balance error, expressed as mm water slice (average over catchment)	mm	mbErrorMm.tss
[2] number of sub-steps needed for channel routing	-	NoSubStepsChannel.tss
[2] number of sub-steps needed for gravity-based soil moisture routine	-	steps.tss

[1] Output only if option 'InitLisflood' = 1 (pre-run)
[2] Output only if option 'InitLisflood' = 0

To speed up the pre-run and to prevent that results are taken from the pre-run, all additional output is disabled if option 'InitLisflood' = 1 is chosen. With 'InitLisflood' = 1 the output is limited to *dis.tss, lzavin.map, lzavin_forest.map* and some other initial maps if additional option like e.g. the double kinematic wave is chosen.

In addition to these time series, by default LISFLOOD reports maps of all state variables at the last timestep of a simulation. These maps can be used to define the initial conditions of a succeeding simulation. For instance, you can do a 1-year simulation on a daily time step, and use the 'end maps' of this simulation to simulate a flood event using an hourly time step. The table below lists all these maps. Note that some state variables are valid for the whole pixel, whereas others are only valid for a sub-domain of each pixel. This is indicated in the last column of the table.

Table: *LISFLOOD default state variable output maps. These maps can be used to define the initial conditions of another simulation.*

Description	Units	File name	Domain
AVERAGE RECHARGE MAP (for lower groundwater zone) (option InitLisflood)			
[1] average inflow to lower zone	$\frac{mm}{timestep}$	lzavin.map	other fraction
[1] average inflow to lower zone (forest)	$\frac{mm}{timestep}$	lzavin_forest.map	forest fraction
INITIAL CONDITION MAPS at defined time steps (option repStateMaps)			

Description	Units	File name	Domain
[2] waterdepth	mm	wdepth00.xxx	whole pixel
[2] channel cross-sectional area	m^2	chcro000.xxx	channel
[2] days since last rain variable	$days$	dslr0000.xxx	other pixel
[2] snow cover zone A	mm	scova000.xxx	snow zone A ($1/3^{rd}$ pixel)
[2] snow cover zone B	mm	scovb000.xxx	snow zone B ($1/3^{rd}$ pixel)
[2] snow cover zone C	mm	scovc000.xxx	snow zone C ($1/3^{rd}$ pixel)
[2] frost index	$\frac{\check{r}C}{days}$	frost000.xxx	other pixel
[2] cumulative interception	mm	cumi0000.xxx	other pixel
[2] soil moisture upper layer	$\frac{mm^3}{mm^3}$	thtop000.xxx	other fraction
[2] soil moisture lower layer	$\frac{mm^3}{mm^3}$	thsub000.xxx	other fraction
[2] water in lower zone	mm	lz000000.xxx	other fraction
[2] water in upper zone	mm	uz000000.xxx	other fraction
[2] days since last rain variable (forest)	$days$	dslF0000.xxx	forest pixel
[2] cumulative interception (forest)	mm	cumF0000.xxx	forest pixel
[2] soil moisture upper layer (forest)	$\frac{mm^3}{mm^3}$	thFt0000.xxx	forest fraction
[2] soil moisture lower layer (forest)	$\frac{mm^3}{mm^3}$	thFs0000.xxx	forest fraction
[2] water in lower zone (forest)	mm	lzF00000.xxx	forest fraction
[2] water in upper zone (forest)	mm	uzF00000.xxx	forest fraction
[2] water in depression storage (sealed)	mm	cseal000.xxx	sealed fraction

[1] Output only if option 'InitLisflood' = 1 (pre-run) [2] Output only if option 'InitLisflood' = 0

Additional output

Apart from the default output, the model can –optionally- generate some additional time series and maps. Roughly this additional output falls in either of the following categories:

Time series

1. Time series with values of model state variables at user-defined locations (sites); reporting of these time series can be activated using the option *repStateSites=1.* Note that 'sites' can be either individual pixels or larger areas (e.g. catchments, administrative areas, and so on). In case of larger areas the model reports the average value for each respective area.
2. Time series with values of model rate variables at user-defined locations (sites); reporting of these time series can be activated using the option *repRateSites=1*
3. Time series with values of meteorological input variables, averaged over the area upstream of each gauge location; reporting of these time series can be activated using the option *repMeteoUpsGauges=1*
4. Time series with values of model state variables, averaged over area upstream of each gauge location; reporting of these time series can be activated using the option *repStateUpsGauges=1*
5. Time series with values of model rate variables, averaged over area upstream of each gauge location; reporting of these time series can be activated using the option *repRateUpsGauges=1*
6. Time series that are specific to other options (e.g. simulation of reservoirs).

Maps

1. Maps of discharge at each time step; reporting of these maps can be activated using the option *repDischargeMaps=1*
2. Maps with values of driving meteorological values at each time step
3. Maps with values of model rate variables at each time step
4. Maps that are specific to other options (e.g. simulation of reservoirs).

In addition, some additional maps and time series may be reported for debugging purposes. In general these are not of any interest to the LISFLOOD user, so they remain undocumented here.

Note that the options *repStateUpsGauges*, *repRateUpsGauges* and *repDischargeMaps* tend to slow down the execution of the model quite dramatically. For applications of the model where performance is critical (e.g. automated calibration runs), we advise to keep them switched off, if possible.

Note again the domains for which variables are valid: all *rate variables* are reported as pixel-average values. Soil moisture and groundwater storage are reported for the permeable fraction of each pixel only. The reported snow cover is the average of the snow depths in snow zones A, B and C.

By default, the names of the reported discharge maps start with the prefix '*dis*' and end with the time step number (the naming conventions are identical to the ones used for the input maps with meteorological variables, which is explained in the annex on LISFLOOD input files. The long table below summarises all options to report additional output maps.

Time series

Table: *LISFLOOD default output time series.*

Description	Units	Settings variable	File name**
RATE VARIABLES AT GAUGES			
[1,2] channel discharge	$\frac{m^3}{s}$	disTS	dis.tss
NUMERICAL CHECKS			
[2] cumulative mass balance error	m^3	WaterMassBalanceTSS	mbError.tss
[2] cumulative mass balance error, expressed as mm water slice (average over catchment)	mm	MassBalanceMMTSS	mbErrorMm.tss
[2] number of sub-steps needed for channel routing	-	NoSubStepsChan	NoSubStepsChannel.tss
[2] number of sub-steps needed for gravity-based soil moisture routine	-	StepsSoilTS	steps.tss

[1] Output only if option 'InitLisflood' = 1 (pre-run) [2] Output only if option 'InitLisflood' = 0

Table: *LISFLOOD optional output time series (only 'InitLisflood' = 0).*

Description	Units	Settings variable	Default name
STATE VARIABLES AT SITES (option *repStateSites*)			
depth of water on soil surface	mm	WaterDepthTS	wDepth.tss
depth of snow cover on soil surface (pixel-average)	mm	SnowCoverTS	snowCover.tss
depth of interception storage	mm	CumInterceptionTS	cumInt.tss
soil moisture content upper layer	$\frac{mm^3}{mm^3}$	Theta1TS	thTop.tss
soil moisture content lower layer	$\frac{mm^3}{mm^3}$	Theta2TS	thSub.tss
storage in upper groundwater zone	mm	UZTS	uz.tss
storage in lower groundwater zone	mm	LZTS	lz.tss
number of days since last rain	$days$	DSLRTS	dslr.tss
frost index	$\frac{\text{ř}C}{days}$	FrostIndexTS	frost.tss
RATE VARIABLES AT SITES (option *repRateSites*)			
rain (excluding snow)	$\frac{mm}{timestep}$	RainTS	rain.tss
Snow	$\frac{mm}{timestep}$	SnowTS	snow.tss
snow melt	$\frac{mm}{timestep}$	SnowmeltTS	snowMelt.tss
actual evaporation	$\frac{mm}{timestep}$	ESActTS	esAct.tss
actual transpiration	$\frac{mm}{timestep}$	TaTS	tAct.tss
rainfall interception	$\frac{mm}{timestep}$	InterceptionTS	interception.tss
evaporation of intercepted water	$\frac{mm}{timestep}$	EWIntTS	ewIntAct.tss
leaf drainage	$\frac{mm}{timestep}$	LeafDrainageTS	leafDrainage.tss
infiltration	$\frac{mm}{timestep}$	InfiltrationTS	infiltration.tss
preferential (bypass) flow	$\frac{mm}{timestep}$	PrefFlowTS	prefFlow.tss
percolation upper to lower soil layer	$\frac{mm}{timestep}$	PercolationTS	dTopToSub.tss
percolation lower soil layer to subsoil	$\frac{mm}{timestep}$	SeepSubToGWTS	dSubToUz.tss
surface runoff	$\frac{mm}{timestep}$	SurfaceRunoffTS	surfaceRunoff.tss

Description	Units	Settings variable	Default name
outflow from upper zone	$\frac{mm}{timestep}$	UZOutflowTS	qUz.tss
outflow from lower zone	$\frac{mm}{timestep}$	LZOutflowTS	qLz.tss
total runoff	$\frac{mm}{timestep}$	TotalRunoffTS	totalRunoff.tss
percolation from upper to lower zone	$\frac{mm}{timestep}$	GwPercUZLZTS	percUZLZ.tss
loss from lower zone	$\frac{mm}{timestep}$	GwLossTS	loss.tss
TIME SERIES, AVERAGE UPSTREAM OF GAUGES			
METEOROLOGICAL INPUT VARIABLES (option *repMeteoUpsGauges*)			
precipitation	$\frac{mm}{timestep}$	PrecipitationAvUpsTS	precipUps.tss
potential reference evapotranspiration	$\frac{mm}{timestep}$	ETRefAvUpsTS	etUps.tss
potential evaporation from soil	$\frac{mm}{timestep}$	ESRefAvUpsTS	esUps.tss
potential open water evaporation	$\frac{mm}{timestep}$	EWRefAvUpsTS	ewUps.tss
average daily temperature	řC	TavgAvUpsTS	tAvgUps.tss
STATE VARIABLES (option *repStateUpsGauges*)			
depth of water on soil surface	mm	WaterDepthAvUpsTS	swdepthUps.tss
depth of snow cover on	mm	SnowCoverAvUpsTS	snowCoverUps.tss
depth of interception storage	mm	CumInterceptionAvUpsTS	cumInterceptionUps.tss
soil moisture upper layer	$\frac{mm^3}{mm^3}$	Theta1AvUpsTS	thTopUps.tss
soil moisture lower layer	$\frac{mm^3}{mm^3}$	Theta2AvUpsTS	thSubUps.tss
groundwater upper zone	mm	UZAvUpsTS	uzUps.tss
groundwater lower zone	mm	LZAvUpsTS	lzUps.tss
number of days since last rain	$days$	DSLRAvUpsTS	dslrUps.tss
frost index	$\frac{\text{ř}C}{days}$	FrostIndexAvUpsTS	frostUps.tss
RATE VARIABLES (option *repRateUpsGauges*)			
rain (excluding snow)	$\frac{mm}{timestep}$	RainAvUpsTS	rainUps.tss
snow	$\frac{mm}{timestep}$	SnowAvUpsTS	snowUps.tss
snow melt	$\frac{mm}{timestep}$	SnowmeltAvUpsTS	snowMeltUps.tss
actual evaporation	$\frac{mm}{timestep}$	ESActAvUpsTS	esActUps.tss
actual transpiration	$\frac{mm}{timestep}$	TaAvUpsTS	tActUps.tss
rainfall interception	$\frac{mm}{timestep}$	InterceptionAvUpsTS	interceptionUps.tss
evaporation of intercepted water	$\frac{mm}{timestep}$	EWIntAvUpsTS	ewIntActUps.tss
leaf drainage	$\frac{mm}{timestep}$	LeafDrainageAvUpsTS	leafDrainageUps.tss
infiltration	$\frac{mm}{timestep}$	InfiltrationAvUpsTS	infiltrationUps.tss
preferential (bypass) flow	$\frac{mm}{timestep}$	PrefFlowAvUpsTS	prefFlowUps.tss
percolation upper to lower soil layer	$\frac{mm}{timestep}$	PercolationAvUpsTS	dTopToSubUps.tss
percolation lower soil layer to subsoil	$\frac{mm}{timestep}$	SeepSubToGWAvUpsTS	dSubToUzUps.tss
surface runoff	$\frac{mm}{timestep}$	SurfaceRunoffAvUpsTS	surfaceRunoffUps.tss
outflow from upper zone	$\frac{mm}{timestep}$	UZOutflowAvUpsTS	qUzUps.tss
outflow from lower zone	$\frac{mm}{timestep}$	LZOutflowAvUpsTS	qLzUps.tss
total runoff	$\frac{mm}{timestep}$	TotalRunoffAvUpsTS	totalRunoffUps.tss
percolation upper to lower zone	$\frac{mm}{timestep}$	GwPercUZLZAvUpsTS	percUZLZUps.tss
loss from lower zone	$\frac{mm}{timestep}$	GwLossTS	lossUps.tss
WATER LEVEL IN CHANNEL (option *repWaterLevelTs*)			
water level in channel	m (above channel bottom)	WaterLevelTS	waterLevel.tss
OUTPUT RELATED TO LOWER ZONE INITIALISATION (option *repLZAvInflowSites* and *repLZAvInflowUpsGauges*)			
average inflow into lower zone	$\frac{mm^3}{day}$	LZAvInflowTS	lzAvIn.tss

Description	Units	Settings variable	Default name
average inflow into lower zone	$\frac{mm^3}{day}$	LZAvInflowAvUpsTS	lzAvInUps.tss

Maps

Table: *LISFLOOD default output maps.*

Description	Units	File name	Domain
AVERAGE RECHARGE MAP (for lower groundwater zone) (option *InitLisflood*)			
[1] average inflow to lower zone	$\frac{mm^3}{day}$	lzavin.map	other fraction
[1] average inflow to lower zone (forest)	$\frac{mm^3}{day}$	lzavin_forest.map	forest fraction
INITIAL CONDITION MAPS at defined time steps (option *repStateMaps*)			
[2] waterdepth	mm	wdepth00.xxx	whole pixel
[2] channel cross-sectional area	m^2	chcro000.xxx	channel
[2] days since last rain variable	$days$	dslr0000.xxx	other pixel
[2] snow cover zone A	mm	scova000.xxx	snow zone A (1/3rd pixel)
[2] snow cover zone B	mm	scovb000.xxx	snow zone B (1/3rd pixel)
[2] snow cover zone C	mm	scovc000.xxx	snow zone C (1/3rd pixel)
[2] frost index	$\frac{\check{r}C}{days}$	frost000.xxx	other pixel
[2] cumulative interception	mm	cumi0000.xxx	other pixel
[2] soil moisture upper layer	$\frac{mm^3}{mm^3}$	thtop000.xxx	other fraction
[2] soil moisture lower layer	$\frac{mm^3}{mm^3}$	thsub000.xxx	other fraction
[2] water in lower zone	mm	lz000000.xxx	other fraction
[2] water in upper zone	mm	uz000000.xxx	other fraction
[2] days since last rain variable (forest)	$days$	dslF0000.xxx	forest pixel
[2] cumulative interception (forest)	mm	cumF0000.xxx	forest pixel
[2] soil moisture upper layer (forest)	$\frac{mm^3}{mm^3}$	thFt0000.xxx	forest fraction
[2] soil moisture lower layer (forest)	$\frac{mm^3}{mm^3}$	thFs0000.xxx	forest fraction
[2] water in lower zone (forest)	mm	lzF00000.xxx	forest fraction
[2] water in upper zone (forest)	mm	uzF00000.xxx	forest fraction
[2] water in depression storage (sealed)	mm	cseal000.xxx	sealed fraction

[1] Output only if option 'InitLisflood' = 1 (pre-run) [2] Output only if option 'InitLisflood' = 0

Table: *LISFLOOD optional output maps (only 'InitLisflood' = 0) .*

Description	Option	Units	Settings variable	Prefix
DISCHARGE AND WATER LEVEL				
discharge	repDischargeMaps	$\frac{m^3}{s}$	DischargeMaps	dis
water level	repWaterLevelMaps	m (above channel bottom)	WaterLevelMaps	wl
METEOROLOGICAL INPUT VARIABLES				
precipitation	repPrecipitationMaps	mm	PrecipitationMaps	pr
potential reference evapotranspiration	repETRefMaps	mm	ETRefMaps	et
potential evaporation from soil	repESRefMaps	mm	ESRefMaps	es
potential open water evaporation	repEWRefMaps	mm	EWRefMaps	ew
average daily temperature	repTavgMaps	mm	TavgMaps	tav

Description	Option	Units	Settings variable	Prefix
STATEVARIABLES				
depth of water on soil surface	repWaterDepthMaps	mm	WaterDepthMaps	wdep
depth of snow cover on soil surface	repSnowCoverMaps	mm	SnowCoverMaps	scov
depth of interception storage	repCumInterceptionMaps	mm	CumInterceptionMaps CumInterceptionForestMaps	cumi cumF
soil moisture content upper layer	repTheta1Maps	$\frac{mm^3}{mm^3}$	Theta1Maps Theta1ForestMaps	thtop thFt
soil moisture content lower layer	repTheta2Maps	$\frac{mm^3}{mm^3}$	Theta2Maps Theta2ForestMaps	thsub thFs
storage in upper groundwater zone	repUZMaps	mm	UZMaps UZForestMaps	uz uzF
storage in lower groundwater zone	repLZMaps	mm	LZMaps LZForestMaps	lz lzF
number of days since last rain	repDSLRMaps	$days$	DSLRMaps DSLRForestMaps	dslr dslF
frost index	repFrostIndexMaps	$\frac{^{\circ}C}{days}$	FrostIndexMaps	frost
RATE VARIABLES				
rain (excluding snow)	repRainMaps	$\frac{mm}{timestep}$	RainMaps	rain
snow	repSnowMaps	$\frac{mm}{timestep}$	SnowMaps	snow
snow melt	repSnowMeltMaps	$\frac{mm}{timestep}$	SnowMeltMaps	smelt
actual evaporation	repESActMaps	$\frac{mm}{timestep}$	ESActMaps	esact
actual transpiration	repTaMaps	$\frac{mm}{timestep}$	TaMaps	tact
rainfall interception	repInterceptionMaps	$\frac{mm}{timestep}$	InterceptionMaps	int
evaporation of intercepted water	repEWIntMaps	$\frac{mm}{timestep}$	EWIntMaps	ewint
leaf drainage	repLeafDrainageMaps	$\frac{mm}{timestep}$	LeafDrainageMaps	ldra
infiltration	repInfiltrationMaps	$\frac{mm}{timestep}$	InfiltrationMaps	inf
preferential (bypass) flow	repPrefFlowMaps	$\frac{mm}{timestep}$	PrefFlowMaps	pflow
percolation upper to lower soil layer	repPercolationMaps	$\frac{mm}{timestep}$	PercolationMaps	to2su
percolation lower soil layer to subsoil	repSeepSubToGWMaps	$\frac{mm}{timestep}$	SeepSubToGWMaps	su2gw
surface runoff	repSurfaceRunoffMaps	$\frac{mm}{timestep}$	SurfaceRunoffMaps	srun
outflow from upper zone	repUZOutflowMaps	$\frac{mm}{timestep}$	UZOutflowMaps	quz
outflow from lower zone	repLZOutflowMaps	$\frac{mm}{timestep}$	LZOutflowMaps	qlz
total runoff	repTotalRunoffMaps	$\frac{mm}{timestep}$	TotalRunoffMaps	trun
percolation upper to lower zone	repGwPercUZLZMaps	$\frac{mm}{timestep}$	GwPercUZLZMaps	uz2lz
loss from lower zone	repGwLossMaps	$\frac{mm}{timestep}$	GwLossMaps	loss

Both the program code and this manual have been carefully inspected before printing. However, no warranties, either expressed or implied, are made concerning the accuracy, completeness, reliability, usability, performance, or fitness for any particular purpose of the information contained in this manual, to the software described in this manual, and to other material supplied in connection therewith. The material is provided "as is". The entire risk as to its quality and performance is with the user.

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