Supporting Materials to "Modular Assessment of Rainfall-Runoff Models Toolbox (MARRMoT) v1.0: an open-source, extendable framework providing implementations of 46 conceptual hydrologic models as continuous space-state formulations"

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S1 Introduction

These Supporting Materials contain documentation for various parts of the MARRMoT software. Section S2 contains model desciptions for the 46 conceptual models included in MARRMoT. Section S3 shows how the constitutive functions of each model are translated into Matlab code, and which models use which of the resulting flux functions. Section S4 shows how 7 different Unit Hydrograph approaches are coded in MARRMoT and which models use these. Section S5 shows an overview of generalized parameter ranges for the 46 models.

S2 Model descriptions

This section contains mathematical descriptions of all models that are included in the Modular Assessment of Rainfall-Runoff Models Toolbox v1.0 (MARRMoT). All descriptions follow the same layout (see the example model at the end of this section):

- Title: gives an informal name for the model structure followed by a unique ID;
- Introduction: gives a brief description of the model, including one or more original reference(s), the number of stores and parameters, a list containing parameter names and occasionally note-worthy deviations from the original model;
- Process list: a brief overview of the main processes the model is intended to represent;
- Figure: a wiring diagram that shows the names of model stores and fluxes;
- Matlab name section: gives the name of the file that contains Matlab code for this model;
- Model equations section: a mathematical description of the model. This uses Ordinary Differential Equations (ODEs) to describe the changes in model storage(s) and constitutive functions that detail how individual fluxes operate.

MARRMoT models intend to stay close to the original models they are based on but differences are unavoidable. We strongly recommend users to read the original paper cited for each model as well as our interpretation given in this document. In many cases, more than one version of a model exists, but these are not always easily distinguishable. There is a certain degree of model name equifinility, where a single name is used to refer to various different version of the same base model. A good example is TOPMODEL, of which many variants exist based around the initial concept of topographic indices. MARRMoT models tend to be based on older rather than newer publications for any given model (to stay close to the "intended" model by the original author(s)) but our selection has been pragmatic to achieve greater variety in the available fluxes and model structures in MARRMoT. The description of each model lists the papers that form the basis of the MARRMoT version of that model.

MARRMoT is set up to work with arbitrary user-defined time step sizes for climate input data. For consistency of parameter values across different time step sizes, the internal dynamics of each model are specified using the base units $[mm]$ and $[d]$. The temporal resolution of climate data is converted to $[mm/d]$ within each model, and model output is converted back to the user-specified time step size. Internal fluxes in each MARRMoT model use the base units and are in $[mm/d]$ and parameter values are specified in the base or derived units (e.g. $[d^{-1}]$ for time coefficients). These units are kept throughout this document.

The computational implementation of constitutive functions is given in section S3 and Unit Hydrographs are specified in section S4. Generalized parameter ranges for all models are given in section S5.

Example model (model ID: nn)

The Example model (fig. S1) is used in the MARRMoT User Manual to show how to create a new MARRMoT model from scratch (Knoben et al., 2018). It has 3 stores and 7 parameters (UZ_{max}, c_{rate}, p_{rate}, k_{lz}, α, k_g, d). The model aims to represent:

- Saturation excess from the upper zone;
- Two-way interaction between upper and lower zone through percolation and capillary rise;
- A split between fast subsurface flow and groundwater recharge from the lower zone;
- Slow runoff from the groundwater;
- Triangular routing of combined surface and subsurface flows.

File names

Model: m_nn_example_7p_3s
Parameter ranges: m_nn_example_7p_3s_parameter ranges

Model equations

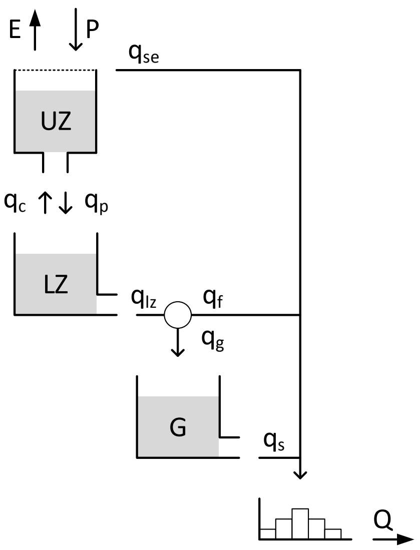

Figure S1: Structure of the Example model

$$\frac{dUZ}{dt} = P + q_c - E - q_{se} - q_p \quad (1)$$

$$E = E_p * \frac{UZ}{UZ_{max}} \quad (2)$$

$$q_c = c_{rate}\left(1 - \frac{UZ}{UZ_{max}}\right) \quad (3)$$

$$q_{se} = \begin{cases} P, & \text{if } UZ = UZ_{max} \\ 0, & \text{otherwise} \end{cases} \quad (4)$$

$$q_p = p_{rate} \quad (5)$$

Where UZ [mm] is the current storage in the upper zone, refilled by precipitation P [mm/d] and capillary rise q_c [mm/d] and drained by evaporation E [mm/d], percolation q_p [mm/d] and saturation excess q_{se} [mm/d]. Evaporation occurs at the potential rate E_p scaled by the current storage in UZ compared to maximum storage UZ_{max} [mm]. Capillary rise occurs at a maximum rate c_{rate} [mm/d] if $UZ = 0$ and decreases linearly if not. Saturation excess flow only occurs when UZ is at maximum capacity.

Percolation occurs at a constant rate p_{rate} $[mm/d]$.

$$\frac{dLZ}{dt} = q_p - q_c - q_{lz} \tag{6}$$
$$q_{lz} = k_{lz} * LZ \tag{7}$$
$$\tag{8}$$

Where LZ [mm] is the current storage in the lower zone, refilled by percolation q_p $[mm/d]$ and drained by capillary rise q_c $[mm/d]$ and outflow q_{lz} $[mm/d]$. Outflow has a linear relation with storage through time parameter k_{lz} $[d^{-1}]$.

$$\frac{dG}{dt} = q_g - q_s \tag{9}$$
$$q_g = \alpha * q_{lz} \tag{10}$$
$$q_s = k_g * G \tag{11}$$

Where G [mm] is the current groundwater storage, refilled by recharge q_g $[mm/d]$ and drained by slow flow q_s $[mm/d]$. Recharge is a fraction α [-] of outflow from the lower zone. Outflow has a linear relation with storage through time parameter k_g $[d^{-1}]$. Saturation excess q_{se}, interflow q_f and slow flow q_s are combined and routed with a triangular Unit Hydrograph with time base d [d] to give outflow Q $[mm/d]$.

S2.1 Collie River Basin 1 (model ID: 01)

The Collie River Basin 1 model (fig. S2) is part of a top-down modelling exercise and is originally applied at the annual scale (Jothityangkoon et al., 2001). This is a classic bucket model. It has 1 store and 1 parameter (S_{max}). The model aims to represent:

- Evaporation from soil moisture;
- Saturation excess surface runoff.

S2.1.1 File names

Model:	m_01_collie1_1s_1p
Parameter ranges:	m_01_collie1_1s_1p_parameter_ ranges

S2.1.2 Model equations

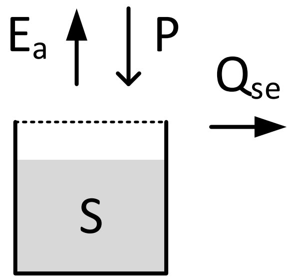

Figure S2: Structure of the Collie River Basin 1 model

$$\frac{dS}{dt} = P - E_a - Q_{se} \tag{12}$$

$$Ea = \frac{S}{S_{max}} * Ep \tag{13}$$

$$Q_{se} = \begin{cases} P, & if\ S > S_{max} \\ 0, & otherwise \end{cases} \tag{14}$$

Where S [mm] is the current storage in the soil moisture and P the precipitation input $[mm/d]$. Actual evaporation E_a $[mm/d]$ is estimated based on the current storage S, the maximum soil moisture storage S_{max} [mm], and the potential evapotranspiration E_p $[mm/d]$. Q_{se} $[mm/d]$ is saturation excess overland flow.

S2.2 Wetland model (model ID: 02)

The Wetland model (fig. S3) is a conceptualization of the perceived dominant processes in a typical Western European wetland (Savenije, 2010). It belongs to a 3-part topography driven modelling exercise, together with a hillslope and plateau conceptualization. Each model is provided in isolation here, because they are well-suited for isolating specific model structure choices. It has 1 store and 4 parameters (D_w, $S_{w,max}$, β_w and K_w). The model aims to represent:

- Stylized interception by vegetation;
- Evaporation;
- Saturation excess runoff generated from a distribution of soil depths;
- A linear relation between storage and slow runoff.

S2.2.1 File names

Model: m_02_wetland_4p_1s
Parameter ranges: m_02_wetland_4p_1s_parameter_ ranges

S2.2.2 Model equations

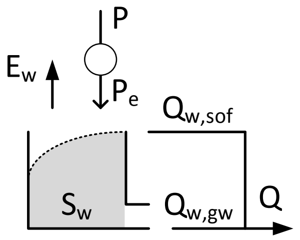

Figure S3: Structure of the Wetland model

$$\frac{dS_w}{dt} = P_e - E_w - Q_{w,sof} - Q_{w,gw} \tag{15}$$

$$P_e = max(P - D_w, 0) \tag{16}$$

$$E_w = \begin{cases} E_p, & \text{if } S_w > 0 \\ 0, & \text{otherwise} \end{cases} \tag{17}$$

$$Q_{w,sof} = \left(1 - \left(1 - \frac{S_w}{S_{w,max}}\right)^{\beta_w}\right) * P_e \tag{18}$$

$$Q_{w,gw} = K_w * S_w \tag{19}$$

Where S_w is the current soil water storage [mm]. Incoming precipitation P [mm/d] is reduced by interception D_w [mm/d], which is assumed to evaporate before the next precipitation event. Evaporation from soil moisture E_w [mm/d] occurs at the potential rate E_p whenever possible. Saturation excess surface runoff $Q_{w,sof}$ [mm/d] depends on the fraction of the catchment that is currently saturated, expressed through parameters $S_{w,max}$ [mm] and β_w [-]. Groundwater flow $Q_{w,gw}$ [mm/d] depends linearly on current storage S_w through parameter K_w [d^{-1}]. Total flow:

$$Q = Q_{w,sof} + Q_{w,gw} \tag{20}$$

S2.3 Collie River Basin 2 (model ID: 03)

The Collie River Basin 2 model (fig. S4) is part of a top-down modelling exercise and is originally applied at the monthly scale (Jothityangkoon et al., 2001). It has 1 store and 4 parameters (S_{max}, S_{fc}, a, M). The model aims to represent:

- Separate bare soil and vegetation evaporation;
- Saturation excess surface runoff;
- Subsurface runoff.

S2.3.1 File names

Model: m_03_collie2_4p_2s
Parameter ranges: m_03_collie2_4p_2s_parameter_ ranges

S2.3.2 Model equations

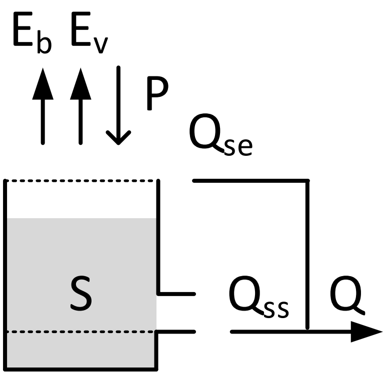

Figure S4: Structure of the Collie River Basin 2 model

$$\frac{dS}{dt} = P - E_b - E_v - Q_{se} - Q_{ss} \tag{21}$$

$$Eb = \frac{S}{S_{max}}(1-M) * Ep \tag{22}$$

$$Ev = \begin{cases} M * E_p, & if\ S > S_{fc} \\ \frac{S}{S_{fc}} * M * E_p, & otherwise \end{cases} \tag{23}$$

$$Q_{se} = \begin{cases} P, & if\ S > S_{max} \\ 0, & otherwise \end{cases} \tag{24}$$

$$Q_{ss} = \begin{cases} a * (S - Sfc), & if\ S > S_{fc} \\ 0, & otherwise \end{cases} \tag{25}$$

Where S [mm] is the current storage in the soil moisture and P $[mm/d]$ the precipitation input. Actual evaporation is split between bare soil evaporation E_b $[mm/d]$ and transpiration through vegetation E_v $[mm/d]$, controlled through the forest fraction M [-]. The evaporation estimates are based on the current storage S, the potential evapotranspiration E_p $[mm/d]$, maximum soil moisture storage S_{max} [mm] and field capacity S_{fc} [mm] respectively. Q_{se} $[mm/d]$ is saturation excess overland flow. Q_{ss} $[mm/d]$ is subsurface flow regulated by runoff coefficient a $[d^{-1}]$. Total flow:

$$Q = Q_{se} + Q_{ss} \tag{26}$$

S2.4 New Zealand model v1 (model ID: 04)

The New Zealand model v1 (fig. S5) is part of a top-down modelling exercise that focusses on several catchments in New Zealand (Atkinson et al., 2002). It has 1 store and 6 parameters (S_{max}, S_{fc}, M, a, b and $t_{c,bf}$). The model aims to represent:

- Separate vegetation and bare soil evaporation;
- Saturation excess overland flow;
- Subsurface runoff when soil moisture exceeds field capacity;
- Baseflow.

S2.4.1 File names

Model: m_04_newzealand1_6p_1s
Parameter ranges: m_04_newzealand1_6p_1s_parameter_ ranges

S2.4.2 Model equations

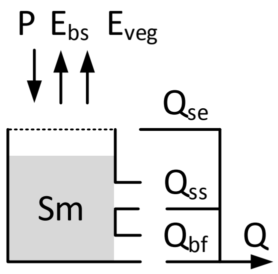

Figure S5: Structure of the New Zealand model v1

$$\frac{dS_m}{dt} = P - E_{veg} - E_{bs} - Q_{se} - Q_{ss} - Q_{bf} \tag{27}$$

$$E_{veg} = \begin{cases} M * E_p, & \text{if } S > S_{fc} \\ \frac{S_m}{S_{fc}} * M * E_p, & \text{otherwise} \end{cases} \tag{28}$$

$$E_{bs} = \frac{S}{S_{max}}(1 - M) * E_p \tag{29}$$

$$Q_{se} = \begin{cases} P, & \text{if } S \geq S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{30}$$

$$Q_{ss} = \begin{cases} \left(a * (S - S_{fc})\right)^b, & \text{if } S \geq S_{fc} \\ 0, & \text{otherwise} \end{cases} \tag{31}$$

$$Q_{bf} = t_{c,bf} * S \tag{32}$$

Where S_m [mm] is the current soil moisture storage which gets replenished through precipitation P $[mm/d]$. Evaporation through vegetation E_{veg} $[mm/d]$ depens on the forest fraction M [-] and field capacity S_{fc} [-]. E_{bs} $[mm/d]$ represents bare soil evaporation. When S exceeds the maximum storage S_{max} [mm], water leaves the model as saturation excess runoff Q_{se}. If S exceeds field capacity S_{fc} [mm], subsurface runoff Q_{ss} $[mm/d]$ is generated controlled by time parameter a $[d^{-1}]$ and nonlinearity parameter b [-]. Q_{bf} represents baseflow controlled by time scale parameter $t_{c,bf}$ $[d^{-1}]$. Total runoff Q_t $[mm/d]$ is:

$$Q_t = Q_{se} + Q_{ss} + Q_{bf} \tag{33}$$

S2.5 IHACRES (model ID: 05)

The IHACRES model (fig. S6) as implemented here is a modification of the original equations (Littlewood et al., 1997; Ye et al., 1997; Croke and Jakeman, 2004), which explicitly account for the various fluxes in a step-wise order. Furthermore, IHACRES usually uses temperature as a proxy for potential evapotranspiration (E_p). Here it uses estimated E_p directly to be consistent with other models. The equations for E_a and U are set up following Croke and Jakeman (2004), with the non-linearity in U based on Ye et al. (1997). THis version thus uses a catchment moisture deficit formulation, rather than a catchment wetness index. Littlewood et al. (1997) recommends the two parallel routing functions. The model has 1 *deficit* store and 6 parameters (lp, d, p, α, τ_q, τ_s). The model aims to represent:

- Catchment deficit build-up
- Slow and fast routing of effective precipitation.

S2.5.1 File names

Model:	m_05_ihacres_6p_1s
Parameter ranges:	m_05_ihacres_6p_1s_parameter_ ranges

S2.5.2 Model equations

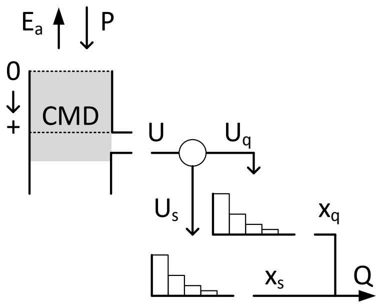

Figure S6: Structure of the IHACRES model

$$\frac{dCMD}{dt} = -P + E_a + U \tag{34}$$

$$E_a = E_p * min\left(1, e^{2\left(1-\frac{CMD}{lp}\right)}\right) \tag{35}$$

$$U = P\left(1 - min\left(1, \left(\frac{CMD}{d}\right)^p\right)\right) \tag{36}$$

$$U_q = \alpha * U \tag{37}$$

$$U_s = (1-\alpha) * U \tag{38}$$

Where CMD is the current moisture deficit [mm], P $[mm/d]$ the incoming precipitation that *reduces* the deficit, E_a $[mm/d]$ evaporation that *increases* the deficit, and U $[mm/d]$ the effective precipitation that occurs when the deficit is below a threshold d [mm].

Evaporation occurs at the potential rate E_p until the moisture deficit reaches wiliting point lp [mm], after which evaporation decreases exponentially with increasing deficit. Effective precipitation U equals incoming precipitation P when the deficit is zero, and decreases as a linear fraction of P until moisture deficit is larger than a threshold d [mm], after which precipitation does not contribute to streamflow any longer. U is divided between fast and slow routing components based on fraction α

[-]. Both routing schemes are exponentially decreasing over time with lags τ_q [d] and τ_s [d] respectively. Total flow:

$$Q = x_q + x_s \tag{39}$$

S2.6 Alpine model v1 (model ID: 06)

The Alpine model v1 model (fig. S7) is part of a top-down modelling exercise and represents a monthly water balance model (Eder et al., 2003). It has 2 stores and 4 parameters (T_t, ddf, S_{max}, t_c). The model aims to represent:

- Snow accumulation and melt;
- Saturation excess overland flow;
- Linear subsurface runoff.

S2.6.1 File names

Model: m_06_alpine1_4p_2s
Parameter ranges: m_06_alpine1_4p_2s_parameter_ ranges

S2.6.2 Model equations

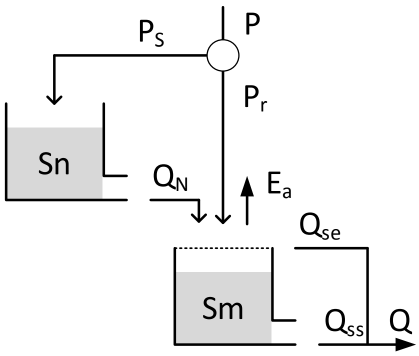

Figure S7: Structure of the Alpine model v1

$$\frac{dSn}{dt} = P_s - Q_N \tag{40}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_t \\ 0, & \text{otherwise} \end{cases} \tag{41}$$

$$Q_N = \begin{cases} ddf * (T - T_t), & \text{if } T \geq T_t \\ 0, & \text{otherwise} \end{cases} \tag{42}$$

Where S_N is the current snow storage [mm], P_s the precipitation that falls as snow $[mm/d]$, Q_N snow melt $[mm/d]$ based on a degree-day factor (ddf, [mm/°C/d]) and threshold temperature for snowfall and snowmelt (T_t, [°C]).

$$\frac{dS_m}{dt} = P_r + Q_N - E_a - Q_{se} - Q_{ss} \tag{43}$$

$$P_r = \begin{cases} P, & \text{if } T > T_t \\ 0, & \text{otherwise} \end{cases} \tag{44}$$

$$E_a = \begin{cases} E_p, & \text{if } S > 0 \\ 0, & \text{otherwise} \end{cases} \tag{45}$$

$$Q_{se} = \begin{cases} P_r + Q_N, & \text{if } S_m \geq S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{46}$$

$$Q_{ss} = t_c * S_m \tag{47}$$

Where S_m [mm] is the current soil moisture storage, which is assumed to evaporate at the potential rate E_p $[mm/d]$ when possible. When S_m exceeds the maximum storage S_{max} [mm], water leaves the model as saturation excess runoff Q_{se}. Q_{ss} represents subsurface flow controlled by time scale parameter t_c $[d^{-1}]$. Total runoff Q_t $[mm/d]$ is:

$$Q_t = Q_{se} + Q_{ss} \tag{48}$$

S2.7 GR4J (model ID: 07)

The GR4J model (fig. S8) is originally developed with an explicit (operator-splitting) time-stepping scheme (Perrin et al., 2003). Recently a new version has been released that works with an implicit time-stepping scheme (Santos et al., 2017). The implementation given here follows most of the equations from Santos et al. (2017), but uses the original Unit Hydrographs for flood routing given by Perrin et al. (2003). It has 2 stores and 4 parameters (x_1, x_2, x_3, x_4). The model aims to represent:

- Implicit interception by vegetation, expressed as net precipitation or evaporation;
- Different time delays within the catchment expressed by two hydrographs;
- Water exchange with neighbouring catchments.

S2.7.1 File names

Model: m_07_gr4j_4p_2s
Parameter ranges: m_07_gr4j_4p_2s_parameter_ ranges

S2.7.2 Model equations

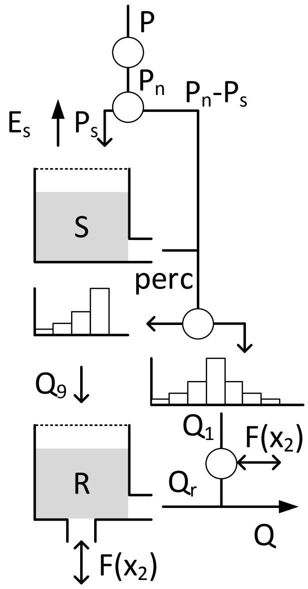

Figure S8: Structure of the GR4J model

$$\frac{dS}{dt} = P_s - E_s - Perc \quad (49)$$

$$P_s = P_n * \left(1 - \left(\frac{S}{x1}\right)^2\right) \quad (50)$$

$$P_n = \begin{cases} P - Ep, & \text{if } P \geq Ep \\ 0, & \text{otherwise} \end{cases} \quad (51)$$

$$E_s = E_n * \left(2\frac{S}{x1} - \left(\frac{S}{x1}\right)^2\right) \quad (52)$$

$$E_n = \begin{cases} Ep - P, & \text{if } Ep > P \\ 0, & \text{otherwise} \end{cases} \quad (53)$$

$$Perc = \frac{x_1^{-4}}{4d} * \left(\frac{4}{9}\right)^{-4} S^5 \quad (54)$$

Where S is the current soil moisture storage [mm], P_s $[mm/d]$ is the fraction of net precipitation P_n $[mm/d]$ redirected to soil moisture, E_s $[mm/d]$ is the fraction of net evaporation E_n $[mm/d]$ subtracted from soil moisture, and $perc$ $[mm/d]$ is percolation to deeper soil layers. Parameter x_1 [mm] is the maximum soil moisture storage.

Percolation *perc* and excess precipitation $P_n - P_s$ are divided into 90% groundwater flow, routed through a triangular routing scheme with time base x_4 [d], and 10% direct runoff, routed through a triangular routing scheme with time base $2x_4$ [d].

$$\frac{dR}{dt} = Q_9 + F(x_2) - Q_r \tag{55}$$

$$F(x_2) = x_2 * \left(\frac{R}{x_3}\right)^{3.5} \tag{56}$$

$$Q_r = \frac{x_3^{-4}}{4d} R^5 \tag{57}$$

Where R [mm] is the current storage in the routing store, $F(x_2)$ $[mm/d]$ the catchment groundwater exchange, depending on exchange coefficient x_2 $[mm/d]$ and the maximum routing capacity x_3 [mm], and Q_r $[mm/d]$ routed flow. Total runoff Q_t $[mm/d]$:

$$Q_t = Q_r + max(Q_1 + F(x_2), 0) \tag{58}$$

S2.8 United States model (model ID: 08)

The United States model (fig. S9) is part of a multi-model comparison study using several catchments in the United States (Bai et al., 2009). It has 2 stores and 5 parameters (α_{ei}, M, S_{max}, fc, α_{ss}). The model aims to represent:

- Interception as a percentage of precipitation;
- Separate unsaturated and saturated zones;
- Separate bare soil evaporation and vegetation transpiration;
- Saturation excess overland flow;
- Subsurface flow.

S2.8.1 File names

Model: m_08_us1_5p_2s
Parameter ranges: m_08_us1_5p_2s_parameter_ ranges

S2.8.2 Model equations

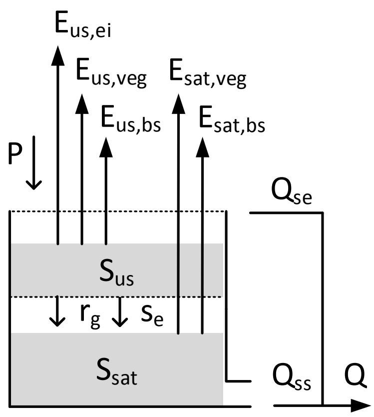

Figure S9: Structure of the United States model

$$\frac{dS_{us}}{dt} = P - E_{us,ei} - E_{us,veg} - E_{us,bs} - r_g \quad (59)$$

$$E_{us,ei} = \alpha_{ei} * P \quad (60)$$

$$E_{us,veg} = \begin{cases} \frac{S_{us}}{S_{us}+S_{sat}} * M * E_p, & \text{if } S_{us} > S_{usfc} \\ \frac{S_{us}}{S_{us}+S_{sat}} * M * E_p * \frac{S_{us}}{S_{usfc}}, & \text{otherwise} \end{cases} \quad (61)$$

$$E_{us,bs} = \frac{S_{us}}{S_{us}+S_{sat}} * (1-M) * \frac{S_{us}}{S_{max}-S_{sat}} * E_p \quad (62)$$

$$r_g = \begin{cases} P, & \text{if } S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \quad (63)$$

$$S_e = \begin{cases} S_{us} - S_{usfc}, & if\ S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \quad (64)$$

$$S_{usfc} = fc * (S_{max} - S_{sat}) \quad (65)$$

Where S_{us} [mm] is the current storage in the unsaturated zone, $E_{us,ei}$ $[mm/d]$ evaporation from interception, $E_{us,veg}$ $[mm/d]$ transpiration through vegetation, $E_{us,bs}$ $[mm/d]$ bare soil evaporation and r_g $[mm/d]$ drainage to the saturated zone. Interception evaporation relies on parameter α_{ei} [-], representing the fraction of precipitation P that is intercepted. The implicit assumption is that this evaporates before the next precipitation event. Transpiration uses forest fraction M [-], potential evapotranspiration E_p $[mm/d]$ and the estimated field capacity S_{usfc} through parameter fc [-]. Bare soil evaporation relies also on the maximum soil moisture storage S_{max} [mm].

$$\frac{dS_{sat}}{dt} = r_g - E_{sat,veg} - E_{sat,bs} - Q_{se} - Q_{ss} \tag{66}$$

$$E_{sat,veg} = \frac{S_{sat}}{S_{max}} * M * E_p \tag{67}$$

$$E_{sat,bs} = \frac{S_{sat}}{S_{max}} * (1 - M) * E_p \tag{68}$$

$$Q_{se} = \begin{cases} r_g, & \text{if } S_{us} \geq S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{69}$$

$$Q_{ss} = \alpha_{ss} * S_{sat} \tag{70}$$

Where S_{sat} [mm] is the current storage in the saturated zone, $E_{sat,veg}$ $[mm/d]$ transpiration through vegetation, $E_{sat,bs}$ $[mm/d]$ bare soil evaporation, Q_{se} $[mm/d]$ saturation excess overland flow and Q_{ss} $[mm/d]$ subsurface flow. Subsurface flow uses time parameter α_{ss} $[d^{-1}]$ Total flow:

$$Q = Q_{se} + Q_{ss} \tag{71}$$

S2.9 Susannah Brook model v1-5 (model ID: 09)

The Susannah Brook model v1-5 (fig. S10) is part of a top-down modelling exercise designed to use auxiliary data (Son and Sivapalan, 2007). It has 2 stores and 6 parameters (S_b, S_{fc}, M, a, b and r). The model aims to represent:

- Evaporation from soil and transpiration from vegetation;
- Saturation excess and non-linear subsurface flow;
- Groundwater recharge and baseflow.

S2.9.1 File names

Model:	m_09_susannah1_6p_2s
Parameter ranges:	m_09_susannah1_6p_2s_parameter_ ranges

S2.9.2 Model equations

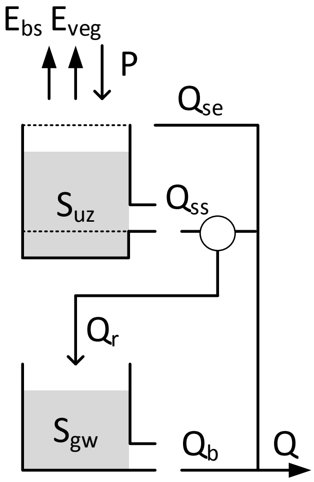

Figure S10: Structure of the Susannah Brook model v1-5

$$\frac{dS_{uz}}{dt} = P - E_{bs} - E_{veg} - Q_{se} - Q_{ss} \tag{72}$$

$$E_{bs} = \frac{S}{S_b}(1 - M)E_p \tag{73}$$

$$E_{veg} = \begin{cases} M * E_p, & \text{if } S > S_{fc} \\ \frac{S}{S_{fc}} M * E_p, & \text{otherwise} \end{cases} \tag{74}$$

$$Q_{se} = \begin{cases} P, & \text{if } S \geq S_b \\ 0, & \text{otherwise} \end{cases} \tag{75}$$

$$Q_{ss} = \begin{cases} \left(\frac{S - S_{fc}}{a}\right)^{\frac{1}{b}}, & \text{if } S > S_{fc} \\ 0, & \text{otherwise} \end{cases} \tag{76}$$

Where S_{uz} is current storage in the upper zone [mm]. P [mm/d] is the precipitation input. E_{bs} is bare soil evaporation [mm/d] based on soil depth S_b [mm] and forest fraction M [-]. E_{veg} is transpiration from vegetation, using the wilting point S_{fc} [mm] and forest fraction M. Q_{se} is saturation excess flow [mm/d]. Q_{ss} is non-linear subsurface flow, using the wilting point S_{fc} as a threshold for flow generation and two flow parameters a [d] and b [-]. Q_r is groundwater recharge [mm/d].

$$\frac{DS_{gw}}{dt} = Q_r - Q_b \tag{77}$$

$$Q_r = r * Q_{ss} \tag{78}$$

$$Q_b = \left(\frac{1}{a} S_{gw}\right)^{\frac{1}{b}} \tag{79}$$

Where S_{gw} is the groundwater storage [mm], and Q_b the baseflow flux [mm/d]. Total flow [mm]:

$$Q = Q_{se} + (Q_{ss} - Q_r) + Q_b \tag{80}$$

S2.10 Susannah Brook model v2 (model ID: 10)

The Susannah Brook model v2 model (fig. S11) is part of a top-down modelling exercise designed to use auxiliary data (Son and Sivapalan, 2007). It has 2 stores and 6 parameters (S_b, ϕ, fc, r, c, d). For consistency with other model formulations, S_b is is used as a parameter, instead of being broken down into its constitutive parts D and ϕ. The model aims to represent:

- Separation of saturated zone and a variable-size unsaturated zone;
- Evaporation from unsaturated and saturated zones;
- Saturation excess and non-linear subsurface flow;
- Deep groundwater recharge.

S2.10.1 File names

Model: m_10_susannah2_6p_2s
Parameter ranges: m_10_susannah2_6p_2s_parameter_ ranges

S2.10.2 Model equations

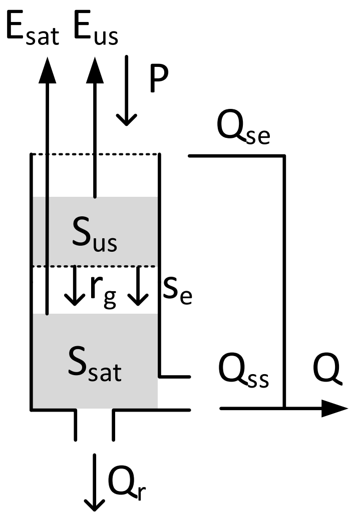

Figure S11: Structure of the Susannah Brook v2 model

$$\frac{dS_{us}}{dt} = P - E_{us} - r_g - S_e \quad (81)$$

$$E_{us} = \frac{S_{us}}{S_b} * E_p \quad (82)$$

$$S_b = D * \phi \quad (83)$$

$$r_g = \begin{cases} P, & if\ S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \quad (84)$$

$$S_e = \begin{cases} S_{us} - S_{usfc}, & if\ S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \quad (85)$$

$$S_{usfc} = (S_b - S_{sat}) * \frac{fc}{\phi} \quad (86)$$

Where S_{us} is the current storage in the unsaturated store [mm], P the current precipitation [mm], S_b [mm] the maximum storage of the soil profile, based on the soil depth D [mm] and the porosity ϕ [-]. r_g is drainage from the unsaturated store to the saturated store [mm], based on the variable field capacity S_{usfc} [mm]. S_{usfc} is based on the current storage on the saturated zone S_{sat} [mm], the maximum soil moisture storage S_b [mm], the field capacity fc [-] and the porosity ϕ [-]. S_e [mm]

is the storage excess, resulting from a decrease of S_{usfc} that leads to more water being stored in the unsaturated zone than should be possible.

$$\frac{dS_{sat}}{dt} = r_g - E_{sat} - Q_{SE} - Q_{SS} - Q_R \tag{87}$$

$$E_{sat} = \frac{S_{sat}}{S_b} * E_p \tag{88}$$

$$Q_{SE} = \begin{cases} r_g + S_e, & \text{if } S_{sat} > S_b \\ 0, & \text{otherwise} \end{cases} \tag{89}$$

$$Q_{SS} = (1-r) * c * (S_{sat})^d \tag{90}$$

$$Q_R = r * c * (S_{sat})^d \tag{91}$$

Where S_{sat} is the current storage in the saturated zone [mm], E_{sat} is the evaporation from the saturated zone [mm], Q_{SE} saturation excess runoff [mm] that occurs when the saturated zone reaches maximum capacity S_b [mm], Q_{SS} is subsurface flow [mm] and Q_R is recharge of deep groundwater [mm]. Both Q_{SS} and Q_R are based on the dimensionless fraction r and subsurface flow constants c [d^{-1}] and d [-]. Total runoff is the sum of Q_{SE} and Q_{SS}:

$$Q = Q_{SE} + Q_{SS} \tag{92}$$

S2.11 Collie River Basin 3 (model ID: 11)

The Collie River Basin 3 model (fig. S12) is part of a top-down modelling exercise and is originally applied at the daily scale (Jothityangkoon et al., 2001). It has 2 stores and 6 parameters (S_{max}, S_{fc}, a, M, b, λ). The model aims to represent:

- Separate bare soil and vegetation evaporation;
- Saturation excess surface runoff;
- Non-linear subsurface runoff;
- Non-linear groundwater runoff.

S2.11.1 File names

Model: m_11_collie3_6p_2s
Parameter ranges: m_11_collie3_6p_2s_parameter_ ranges

S2.11.2 Model equations

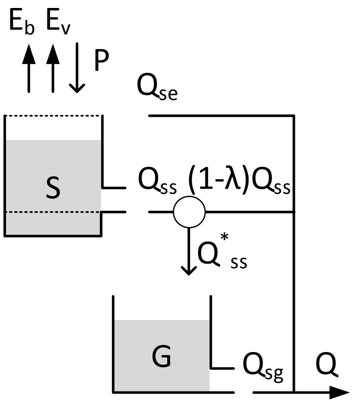

Figure S12: Structure of the Collie River Basin 3 model

$$\frac{dS}{dt} = P - E_b - E_v - Q_{se} - Q_{ss} \tag{93}$$

$$Eb = \frac{S}{S_{max}}(1-M)*Ep \tag{94}$$

$$Ev = \begin{cases} M*E_p, & if\ S > S_{fc} \\ \frac{S}{S_{fc}}*M*E_p, & otherwise \end{cases} \tag{95}$$

$$Q_{se} = \begin{cases} P, & if\ S > S_{max} \\ 0, & otherwise \end{cases} \tag{96}$$

$$Q_{ss} = \begin{cases} \left(a*(S-Sfc)\right)^b, & if\ S > S_{fc} \\ 0, & otherwise \end{cases} \tag{97}$$

Where S [mm] is the current storage in the soil moisture and P the precipitation input $[mm/d]$. Actual evaporation is split between bare soil evaporation E_b $[mm/d]$ and transpiration through vegetation E_v $[mm/d]$, controlled through the forest fraction M. The evaporation estimates are based on the current storage S, the potential evapotranspiration E_p $[mm/d]$ and the maximum soil moisture storage S_{max} [mm], and field capacity S_{fc} [mm] respectively. Q_{se} $[mm/d]$ is saturation excess overland flow. Q_{ss} $[mm/d]$ is non-linear subsurface flow regulated by runoff coefficients a $[d^{-1}]$ and b [-].

$$\frac{dG}{dt} = Q^*_{ss} - Qsg \tag{98}$$

$$Q^*_{ss} = \lambda * Q_{ss} \tag{99}$$

$$Q_{sg} = (a * G)^b \tag{100}$$

Where G [mm] is groundwater storage. Q^*_{ss} $mm/d]$ is the fraction of Q_{ss} directed to groundwater. Q_{sg} $[mm/d]$ is non-linear groundwater flow that relies on the same parameters as subsurface flow uses. Total runoff:

$$Q = Q_{se} + (1 - \lambda) * Q_{ss} + Q_{sg} \tag{101}$$

S2.12 Alpine model v2 (model ID: 12)

The Alpine model v2 (fig. S13) is part of a top-down modelling exercise and represents a daily water balance model (Eder et al., 2003). It has 2 stores and 6 parameters (T_t, ddf, S_{max}, C_{fc}, $t_{c,in}$, $t_{c,bf}$). The model aims to represent:

- Snow accumulation and melt;
- Saturation excess overland flow;
- Linear subsurface runoff.

S2.12.1 File names

Model: m_12_alpine2_6p_2s
Parameter ranges: m_12_alpine2_6p_2s_parameter_ ranges

S2.12.2 Model equations

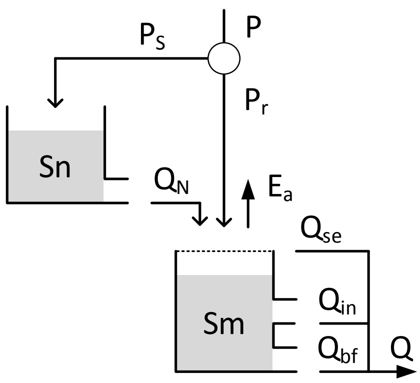

Figure S13: Structure of the Alpine model v1

$$\frac{dSn}{dt} = P_s - Q_N \tag{102}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_t \\ 0, & \text{otherwise} \end{cases} \tag{103}$$

$$Q_N = \begin{cases} ddf * (T - T_t), & \text{if } T \geq T_t \\ 0, & \text{otherwise} \end{cases} \tag{104}$$

Where S_N is the current snow storage [mm], P_s the precipitation that falls as snow $[mm/d]$, Q_N snow melt $[mm/d]$ based on a degree-day factor (ddf, [mm/°C/d]) and threshold temperature for snowfall and snowmelt (T_t, [°C]).

$$\frac{dS}{dt} = P_r + Q_N - E_a - Q_{se} - Q_{in} - Q_{bf} \tag{105}$$

$$P_r = \begin{cases} P, & \text{if } T > TT \\ 0, & \text{otherwise} \end{cases} \tag{106}$$

$$E_a = \begin{cases} E_p, & \text{if } S > 0 \\ 0, & \text{otherwise} \end{cases} \tag{107}$$

$$Q_{se} = \begin{cases} P_r + Q_N, & \text{if } S \geq S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{108}$$

$$Q_{in} = \begin{cases} t_{c,in} * (S - S_{fc}), & \text{if } S > S_{fc} \\ 0, & \text{otherwise} \end{cases} \tag{109}$$

$$Q_{bf} = t_{c,bf} * S \tag{110}$$

Where S [mm] is the current soil moisture storage, which is assumed to evaporate at the potential rate E_p $[mm/d]$ when possible. When S exceeds the maximum storage S_{max} [mm], water leaves the model as saturation excess runoff Q_{se}. If S exceeds field capacity S_{fc} [mm], interflow Q_{in} $[mm/d]$ is generated controlled by time parameter $t_{c,in}$ $[d^{-1}]$. Q_{bf} represents baseflow controlled by time scale parameter $t_{c,bf}$ $[d^{-1}]$. Total runoff Q_t $[mm/d]$ is:

$$Q_t = Q_{se} + Q_{in} + Q_{bf} \tag{111}$$

S2.13 Hillslope model (model ID: 13)

The Hillslope model (fig. S14) is a conceptualization of the perceived dominant processes in a typical Western European hillslope (Savenije, 2010). It belongs to a 3-part topography driven modelling exercise, together with a wetland and plateau conceptualization. Each model is provided in isolation here, because they are well-suited for isolating specific model structure choices. It has 2 store and 7 parameters (D_w, $S_{h,max}$, β_h, a, T_h, C and K_h). The model aims to represent:

- Stylized interception by vegetation;
- Evaporation;
- Separation between rapid subsurface flow and groundwater recharge;
- Capillary rise and linear relation runoff from groundwater.

S2.13.1 File names

Model:	m_13_hillslope_7p_2s
Parameter ranges:	m_13_hillslope_7p_2s_parameter_ ranges

S2.13.2 Model equations

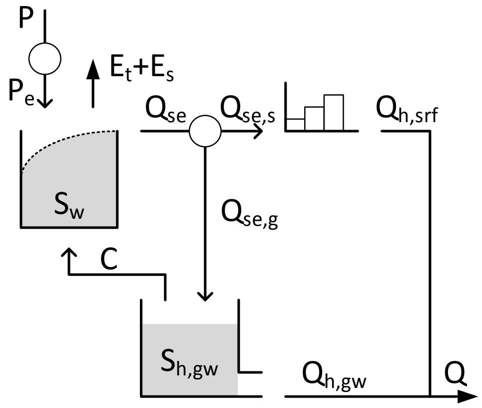

Figure S14: Structure of the Hillslope model

$$\frac{dS_w}{dt} = P_e + C - (E_t + E_s) - Q_{se} \tag{112}$$

$$P_e = max(P - D_h, 0) \tag{113}$$

$$C = c. \tag{114}$$

$$E_t + E_s = \begin{cases} E_p, & \text{if } S_w > 0 \\ 0, & \text{otherwise} \end{cases} \tag{115}$$

$$Q_{se} = \left(1 - \left(1 - \frac{S_h}{S_{h,max}}\right)^{\beta_h}\right) * P_e \tag{116}$$

(117)

Where S_w is the current soil water storage [mm]. Incoming precipitation P [mm/d] is reduced by interception D_h [mm/d], which is assumed to evaporate before the next precipitation event. C is capillary rise from groundwater [mm/d], given as a constant rate. Evaporation from soil moisture $E_t + E_s$ [mm/d] occurs at the potential rate E_p whenever possible. Storage excess surface runoff Q_{se} [mm/d] depends on the fraction of the catchment that is currently saturated, expressed through parameters $S_{h,max}$ [mm] and β_h [-].

$$\frac{dS_{h,gw}}{dt} = Q_{se,g} - C - Q_{h,gw} \tag{118}$$

$$Q_{se,g} = (1 - a) * Q_{se} \tag{119}$$

$$Q_{h,gw} = K_h * S_{h,gw} \tag{120}$$

Where $S_{h,gw}$ is current groundwater storage [mm]. $Q_{se,g}$ is the groundwater fraction of storage excess flow Q_{se} [mm/d], with $Q_{se,s}$ as its complementary part. a is the parameter controlling this division [-]. Groundwater flow $Q_{h,gw}$ [mm/d] depends linearly on current storage $S_{h,gw}$ through parameter K_h [d^{-1}]. Total flow Q_t is the sum of $Q_{h,gw}$ and $Q_{h,srf}$, the latter of which is $Q_{se,s}$ lagged over T_h days.

S2.14 TOPMODEL (model ID: 14)

The TOPMODEL (fig. S15) is originally a semi-distributed model that relies on topographic information (Beven and Kirkby, 1979). The model(ling concept) has undergone many revisions and significant differences can be seen between various publications. The version presented here is mostly based on Beven et al. (1995), with several necessary simplifications. Following Clark et al. (2008), the model is simplified to a lumped model (removing the distributed routing component) and all parameters are calibrated. This means the distribution of topographic index values that characterizes TOPMODEL are estimated using a shifted 2-parameter gamma distribution instead of being based on DEM data (Sivapalan et al., 1987; Clark et al., 2008). For simplicity of the evaporation calculations, the root zone store and unsaturated zone store are combined into a single threshold store with identical functionality to the original 2-store concept. The model has 2 stores and 7 parameters ($S_{UZ,max}$, S_t, K_d, q_0, f, χ, ϕ). The model aims to represent:

- Variable saturated area with direct runoff from the saturated part;
- Infiltration and saturation excess flow;
- Leakage to, and non-linear baseflow from, a deficit store.

S2.14.1 File names

Model: m_14_topmodel_7p_2s
Parameter ranges: m_14_topmodel_7p_2s_parameter_ ranges

S2.14.2 Model equations

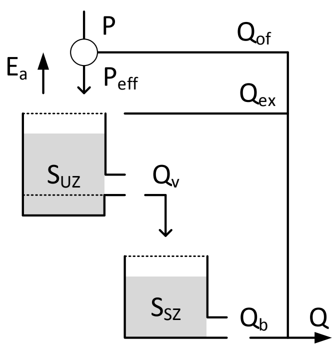

Figure S15: Structure of the TOPMODEL

$$\frac{dS_{UZ}}{dt} = P_{eff} - Q_{ex} - E_a - Q_v \tag{121}$$

$$P_{eff} = P - Q_{of} = P - A_C * P \tag{122}$$

$$Q_{ex} = \begin{cases} P_{eff}, & \text{if } S_{UZ} = S_{uz,max} \\ 0, & \text{otherwise} \end{cases} \tag{123}$$

$$E_a = \begin{cases} E_p, & \text{if } S_{UZ} > S_t * S_{UZ,max} \\ \frac{S_{UZ}}{S_t * S_{UZ,max}} * E_p, & \text{otherwise} \end{cases} \tag{124}$$

$$Q_v = \begin{cases} k_d \frac{S_{UZ} - S_t * S_{UZ,max}}{S_{UZ,max}(1 - S_t)}, & \text{if } S_{UZ} > S_t * S_{UZ,max} \\ 0, & \text{otherwise} \end{cases} \tag{125}$$

Where S_{UZ} [mm] is the current storage in the combined unsaturated zone and root zone, with S_t [-] (fraction of $S_{UZ,max}$) indicating the boundary between the two and being the threshold above which drainage to the saturated zone can occur. P_{eff} $[mm/d]$ is the fraction of precipitation that does not fall on the saturated area A_c [-], E_a $[mm/d]$ is evaporation that occurs at the potential rate for the unsaturated zone and scaled linearly with storage in the root zone, Q_{ex} $[mm/d]$ is overflow when the bucket reaches maximum capacity $S_{UZ,max}$ [mm], and Q_v $[mm/d]$ is drainage to the saturated zone, depending on time parameter k_d $[d^{-1}]$ and the relative storage in the unsaturated zone compared to the current deficit in the saturated zone.

$$\frac{dS_{SZ}}{dt} = -Q_v + Q_b \tag{126}$$

$$Q_b = q_0 * e^{-f*S_{SZ}} \tag{127}$$

Where S_{SZ} [mm] is the current storage *deficit* in the saturated zone store, which is increased by baseflow Q_b $[mm/d]$ and decreased by drainage Q_v. Q_b relies on saturated flow rate q_0 $[mm/d]$, parameter f $[mm^{-1}]$ and current deficit S_{SZ}. Total flow:

$$Q = Q_{of} + Q_{ex} + Q_b \tag{128}$$

$$Q_{of} = A_c * P \tag{129}$$

The saturated area A_c is calculated as follows. First, the within-catchment distribution of topographic index values is estimated with a shifted 2-parameter gamma distribution (Sivapalan et al., 1987; Clark et al., 2008):

$$f(\zeta) = \begin{cases} \frac{1}{\chi\Gamma(\phi)}\left(\frac{\zeta-\mu}{\chi}\right)^{\phi-1} exp\left(-\frac{\zeta-\mu}{\chi}\right), & \text{if } \zeta > \mu \\ 0, & \text{otherwise} \end{cases} \tag{130}$$

Where Γ is the gamma function and χ, ϕ and μ are parameters of the gamma distribution. Following Clark et al. (2008), μ is fixed at $\mu = 3$ and χ and ϕ are calibration parameters. ζ represents the topographic index $ln(a/tan\beta)$ with mean value $\lambda = \chi\phi + \mu$. Saturated area A_c is computed as the fraction of the catchment that is above a deficit-dependent critical value ζ_{crit}:

$$A_c = \int_{\zeta_{crit}}^{\infty} f(\zeta)d\zeta \tag{131}$$

$$\zeta_{crit} = f * S_{SZ} + \lambda \tag{132}$$

S2.15 Plateau model (model ID: 15)

The Plateau model (fig. S16) is a conceptualization of the perceived dominant processes in a typical Western European plateau (Savenije, 2010). It belongs to a 3-part topography driven modelling exercise, together with a wetland and hillslope conceptualization. Each model is provided in isolation here, because they are well-suited for isolating specific model structure choices. It has 2 stores and 8 parameters (F_{max}, D_p, $S_{u,max}$, lp, p, T_p, C and K_p). The model aims to represent:

- Stylized interception by vegetation;
- Evaporation controlled by a wilting point and moisture constrained transpiration;
- Separation between infiltration and infiltration excess flow;
- Capillary rise and linear relation runoff from groundwater.

S2.15.1 File names

Model: m_15_plateau_8p_2s
Parameter ranges: m_15_plateau_8p_2s_parameter_ ranges

S2.15.2 Model equations

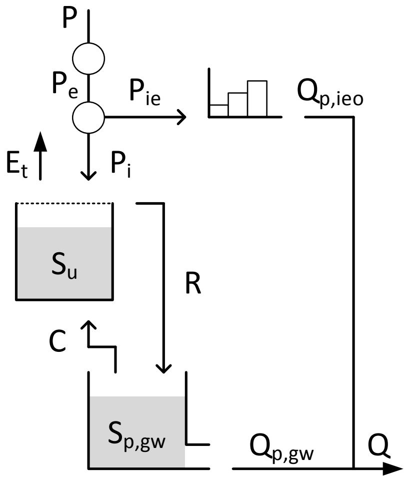

Figure S16: Structure of the Plateau model

$$\frac{dS_u}{dt} = P_i + C - E_t - R \tag{133}$$

$$P_i = min(P_e, F_{max}) \tag{134}$$

$$= min\left(max(P - D_p, 0), F_{max}\right) \tag{135}$$

$$C = c. \tag{136}$$

$$E_t = E_p * max\left(p\frac{S_u - S_{wp}}{S_{u,max} - S_{wp}}, 0\right) \tag{137}$$

$$R = \begin{cases} P_i + C, & \text{if } S_u = S_{u,max} \\ 0, & \text{otherwise} \end{cases} \tag{138}$$

Where S_u is the current soil water storage [mm]. Incoming precipitation P [mm/d] is reduced by interception D_p [mm/d], which is assumed to evaporate before the next precipitation event. P_e is further divided into infiltration P_i [mm/d] based on the maximum infiltration rate F_{max} [mm/d] and infiltration excess $P_{ie} = P_e - P_i$ [mm/d]. C is capillary rise from ground water [mm/d], given as a constant rate.

Evaporation from soil moisture E_t [mm/d] occurs at the potential rate E_p when S_u is above the wilting point S_{wp} [mm] (here defined as $S_{wp} = lp * S_{u,max}$) and is further constrained by coefficient p [-], which is between 0 and 1. Storage excess R [mm/d] flows into the groundwater.

$$\frac{dS_{p,gw}}{dt} = R - C - Q_{p,gw} \tag{139}$$

$$Q_{p,gw} = K_p * S_{p,gw} \tag{140}$$

Where $S_{p,gw}$ is current groundwater storage [mm]. Groundwater flow $Q_{p,gw}$ [mm/d] depends linearly on current storage $S_{p,gw}$ through parameter K_p [d^{-1}]. Total flow Q_t is the sum of $Q_{p,gw}$ and $Q_{p,ieo}$, the latter of which is P_{ie} lagged over T_p days.

S2.16 New Zealand model v2 (model ID: 16)

The New Zealand model v2 (fig. S17) is part of a top-down modelling exercise that focusses on several catchments in New Zealand (Atkinson et al., 2002). It has 2 stores and 8 parameters (I_{max}, S_{max}, S_{fc}, M, a, b and $t_{c,bf}$, d). The model aims to represent:

- Interception by vegetation;
- Separate vegetation and bare soil evaporation;
- Saturation excess overland flow;
- Subsurface runoff when soil moisture exceeds field capacity;
- Baseflow;
- Flow routing.

S2.16.1 File names

Model: m_16_newzealand2_8p_2s
Parameter ranges: m_16_newzealand2_8p_2s_parameter_ ranges

S2.16.2 Model equations

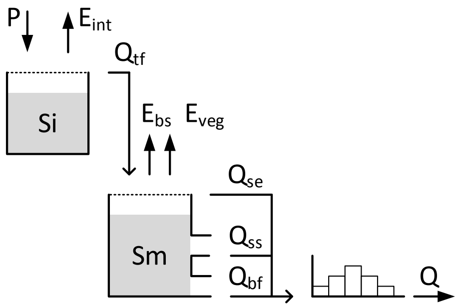

Figure S17: Structure of the New Zealand model v1

$$\frac{dS_i}{dt} = P - E_{int} - Q_{tf} \tag{141}$$

$$E_{int} = Ep \tag{142}$$

$$Q_{tf} = \begin{cases} P, & \text{if } S_i \geq I_{max} \\ 0, & \text{otherwise} \end{cases} \tag{143}$$

Where S_i [mm] is the current interception storage which gets replenished through daily precipitation P [mm/d]. Intercepted water is assumed to evaporate (E_{int} [mm/d]) at the potential rate E_p [mm/d] when possible. Q_{tf} [mm/d] represents throughfall towards soil moisture.

$$\frac{dS_m}{dt} = Q_{tf} - E_{veg} - E_{bs} - Q_{se} - Q_{ss} - Q_{bf} \tag{144}$$

$$E_{veg} = \begin{cases} M * E_p, & \text{if } S > S_{fc} \\ \frac{S_m}{S_{fc}} * M * E_p, & \text{otherwise} \end{cases} \tag{145}$$

$$E_{bs} = \frac{S}{S_{max}}(1 - M) * E_p \tag{146}$$

$$Q_{se} = \begin{cases} P, & \text{if } S \geq S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{147}$$

$$Q_{ss} = \begin{cases} \left(a * (S - S_{fc})\right)^b, & \text{if } S \geq S_{fc} \\ 0, & \text{otherwise} \end{cases} \tag{148}$$

$$Q_{bf} = t_{c,bf} * S \tag{149}$$

Where S_m [mm] is the current soil moisture storage which gets replenished through daily precipitation P $[mm/d]$. Evaporation through vegetation E_{veg} $[mm/d]$ depens on the forest fraction M [-] and field capacity S_{fc} [-]. E_{bs} $[mm/d]$ represents bare soil evaporation. When S exceeds the maximum storage S_{max} [mm], water leaves the model as saturation excess runoff Q_{se}. If S exceeds field capacity S_{fc} [mm], subsurface runoff Q_{ss} $[mm/d]$ is generated controlled by time parameter a $[d^{-1}]$ and nonlinearity parameter b [-]. Q_{bf} represents baseflow controlled by time scale parameter $t_{c,bf}$ $[d^{-1}]$. Total runoff Q_t $[mm/d]$ is:

$$Q_t = Q_{se} + Q_{ss} + Q_{bf} \tag{150}$$

Total flow is delayed by a triangular routing scheme controlled by time parameter d [d].

S2.17 Penman model (model ID: 17)

The Penman model (fig. S18) is based on the drying curve concept described in Penman (1950) (Wagener et al., 2002). It has 3 stores and 4 parameters (S_{max}, ϕ, α, k_1). The model aims to represent:

- Moisture accumulation and evaporation from the root zone;
- Bypass of excess moisture to the stream;
- Deficit-based groundwater accounting;
- Linear flow routing.

S2.17.1 File names

Model: m_17_penman_4p_3s
Parameter ranges: m_17_penman_4p_3s_parameter_ ranges

S2.17.2 Model equations

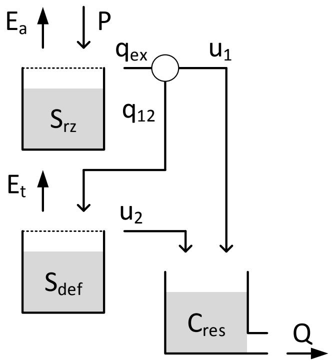

Figure S18: Structure of the Penman model

$$\frac{dS_{rz}}{dt} = P - E_a - Q_{ex} \tag{151}$$

$$E_a = \begin{cases} E_p, & \text{if } S_{rz} > 0 \\ 0, & \text{otherwise} \end{cases} \tag{152}$$

$$P_{ex} = \begin{cases} P, & \text{if } S_{rz} = S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{153}$$

Where S_{rz} [mm] is the current storage in the root zone, refilled by precipitation P $[mm/d]$ and drained by evaporation E_a $[mm/d]$ and moisture excess q_{ex} $[mm/d]$. E_a occurs at the potential rate E_p $[mm/d]$ whenever possible. q_{ex} occurs only when the store is at maximum capacity S_{max} [mm].

$$\frac{dS_{def}}{dt} = E_t + u_2 - q_{12} \tag{154}$$

$$E_t = \begin{cases} \gamma * E_p, & \text{if } S_{rz} = 0 \\ 0, & \text{otherwise} \end{cases} \tag{155}$$

$$u_2 = \begin{cases} q_{12}, & \text{if } S_{def} = 0 \\ 0, & \text{otherwise} \end{cases} \tag{156}$$

$$q_{12} = (1 - \phi) * q_{ex} \tag{157}$$

Where S_{def} [mm] is the current moisture *deficit*, which is increased by evaporation E_t $[mm/d]$ and reduced by inflow q_{12} $[mm/d]$. E_t occurs only when the upper store

S_{rz} is empty and at a fraction γ [-] of E_p. Inflow q_{12} is the fraction $(1-\phi)$ [-] of q_{ex} that does not bypass the lower soil layer. Saturation excess u_2 $[mm/d]$ occurs only when there is zero deficit.

$$\frac{dC_{res}}{dt} = u_1 + u_2 - Q \tag{158}$$

$$Q = k_1 * C_{res} \tag{159}$$

Where C_{res} [mm] is the current storage in the routing reservoir, increased by u_1 and u_2, and drained by runoff Q $[mm/d]$. Q has a linear relationship with storage through time scale parameter k_1 $[d^{-1}]$.

S2.18 SIMHYD (model ID: 18)

The SIMHYD model (fig. S19) is a simplified version of MODHYDROLOG, originally developped for use in Australia (Chiew et al., 2002). It has 3 stores (I, SMS, GW) and 7 parameters (INSC, COEFF, SQ, SMSC, SUB, CRAK, K). The model aims to represent:

- Interception by vegetation;
- Infiltration and infiltration excess flow;
- Preferential groundwater recharge, interflow and saturation excess flow;
- Groundwater recharge resulting from filling up of soil moisture storage capacity;
- Slow flow from groundwater.

S2.18.1 File names

Model: m_18_simhyd_7p_3s
Parameter ranges: m_18_simhyd_7p_3s_parameter_ ranges

S2.18.2 Model equations

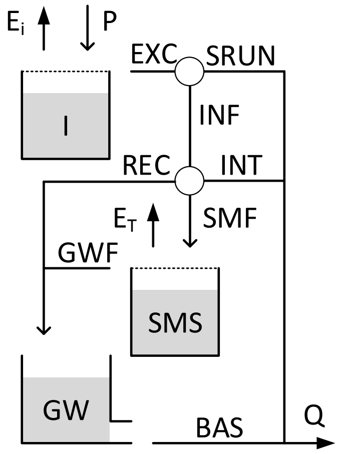

Figure S19: Structure of the SIMHYD model

$$\frac{dI}{dt} = P - E_i - EXC \tag{160}$$

$$E_i = \begin{cases} E_p, & \text{if } I > 0 \\ 0, & \text{otherwise} \end{cases} \tag{161}$$

$$EXC = \begin{cases} P, & \text{if } I = INSC \\ 0, & \text{otherwise} \end{cases} \tag{162}$$

Where I is the current interception storage [mm], P precipitation [mm/d], E_i the evaporation from the interception store [mm/d] and EXC the excess rainfall [mm/d]). Evaporation is assumed to occur at the potential rate when possible. When I exceeds the maximum interception capacity $INSC$ [mm], water is routed to the rest of the model as excess precipitation EXC.

$$\frac{dSMS}{dt} = SMF - E_T - GWF \tag{163}$$
$$SMF = INF - INT - REC \tag{164}$$
$$INF = min\left(COEFF * exp\left(\frac{-SQ * SMS}{SMSC}\right), EXC\right) \tag{165}$$
$$INT = SUB * \frac{SMS}{SMSC} * INF \tag{166}$$
$$REC = CRAK * \frac{SMS}{SMSC} * (INF - INT) \tag{167}$$
$$E_T = min\left(10 * \frac{SMS}{SMSC}, PET\right) \tag{168}$$
$$GWF = \begin{cases} SMF, & \text{if } SMS = SMSC \\ 0, & \text{otherwise} \end{cases} \tag{169}$$

Where SMS is the current storage in the soil moisture store [mm]. INF is total infiltration [mm/d] from excess precipitation, based on maximum infiltration loss parameter COEFF [-], the infiltration loss exponent SQ [-] and the ratio between current soil misture storage SMS and the maximum soil moisture capacity SMSC [mm]. INT represents interflow and saturation excess flow [mm/d], using a constant of proportionality SUB [-]. REC is preferential recharge of groundwater [mm/d] based on another constant of proportionality CRAK [-]. SMF is flow into soil moisture storage [mm/d]. E_T evaporation from the soil moisture that occurs at the potential rate when possible [mm/d], and GWF the flow to the groundwater store [mm/d]:

$$\frac{dGW}{dt} = REC + GWF - BAS \tag{170}$$
$$BAS = K * GW \tag{171}$$

Where GW is the current storage [mm] in the groundwater reservoir. Outflow BAS [mm/d] from the reservoir has a linear relation with storage through the linear recession parameter K [d^{-1}]. Total outflow Q_t [mm/d] is the sum of three parts:

$$Q_t = SRUN + INT + BAS \tag{172}$$
$$SRUN = EXC - INF \tag{173}$$

S2.19 Australia model (model ID: 19)

The Australia model (fig. S20) is part of a top-down modelling exercise designed to use auxiliary data (Farmer et al., 2003). Some adjustments were made to the evaporation equations: these were originally separated between vegetation and bare soil evaporation, scaled between the unsaturated and saturated zone. This has been simplified to separation between unsaturated and saturated evaporation only. The model has 3 stores and 8 parameters (S_b, ϕ, fc, α_{SS}, β_{SS}, K_{deep}, α_{BF}, β_{BF}). For consistency with other model formulations, S_b is is used as a parameter, instead of being broken down into its constitutive parts D and ϕ. The model aims to represent:

- Separation of saturated zone and a variable-size unsaturated zone;
- Evaporation from unsaturated and saturated zones;
- Saturation excess and non-linear subsurface flow;
- Deep groundwater recharge and baseflow.

S2.19.1 File names

Model: m_19_australia_8p_3s
Parameter ranges: m_19_australia_8p_3s_parameter_ ranges

S2.19.2 Model equations

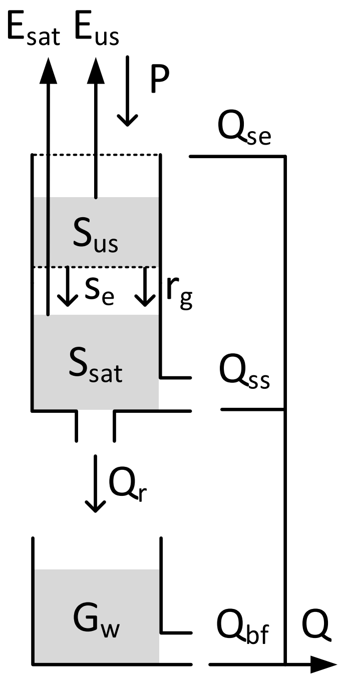

Figure S20: Structure of the Australia model

$$\frac{dS_{us}}{dt} = P - E_{us} - r_g - s_e \tag{174}$$

$$E_{us} = \frac{S_{us}}{S_b} * E_p \tag{175}$$

$$S_b = D * \phi \tag{176}$$

$$r_g = \begin{cases} P, & if\ S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \tag{177}$$

$$s_e = \begin{cases} S_{us} - S_{usfc}, & if\ S_{us} > S_{usfc} \\ 0, & \text{otherwise} \end{cases} \tag{178}$$

$$S_{usfc} = (S_b - S_{sat}) * \frac{fc}{\phi} \tag{179}$$

Where S_{us} is the current storage in the unsaturated store [mm], P the current precipitation [mm/d], S_b [mm] the maximum storage of the soil profile, based on the soil depth D [mm] and the porosity ϕ [-]. r_g [mm/d] is drainage from the unsaturated store to the saturated store, based on the variable field capacity S_{usfc} [mm]. S_{usfc} is based on the current storage on the saturated zone S_{sat} [mm], the maximum soil moisture storage S_b [mm], the field capacity fc [-] and the porosity ϕ [-]. s_e $[mm/d]$ is the storage excess, resulting from a decrease of S_{usfc} that leads to more water being stored in the unsaturated zone than should be possible.

$$\frac{dS_{sat}}{dt} = r_g - E_{sat} - Q_{SE} - Q_{SS} - Q_R \tag{180}$$

$$E_{sat} = \frac{S_{sat}}{S_b} * E_p \tag{181}$$

$$Q_{SE} = \begin{cases} r_g + S_e, & \text{if } S_{sat} > S_b \\ 0, & \text{otherwise} \end{cases} \tag{182}$$

$$Q_{SS} = \alpha_{SS} * (S_{sat})^{\beta_{SS}} \tag{183}$$

$$Q_R = K_{deep} * S_{sat} \tag{184}$$

Where S_{sat} is the current storage in the saturated zone [mm], E_{sat} is the evaporation from the saturated zone [mm], Q_{SE} saturation excess runoff $[mm/d]$ that occurs when the saturated zone reaches maximum capacity S_b [mm], Q_{SS} is subsurface flow $[mm/d]$ and Q_R is recharge of deep groundwater $[mm/d]$. Both Q_{SS} and Q_R are based on the dimensionless fraction r and subsurface flow constants c $[d^{-1}]$ and d [-].

$$\frac{dG_w}{dt} = Q_R - Q_{BF} \tag{185}$$

$$Q_{BF} = \alpha_{BF} * (G_w)^{\beta_{BF}} \tag{186}$$

$$\tag{187}$$

Where G_w is the current groundwater storage [mm] and Q_{BF} baseflow, dependent on parameters α_{BF} $[d^{-1}]$ and β_{BF} [-]. Total runoff is the sum of QSE, Q_{SS} and Q_{BF}:

$$Q = Q_{SE} + Q_{SS} + Q_{BF} \tag{188}$$

S2.20 Generalized Surface inFiltration Baseflow model (model ID: 20)

The GSFB model (fig. S21) is originally developed for use in Australian ephemeral catchments (Nathan and McMahon, 1990; Ye et al., 1997). It has 3 stores and 8 parameters (C. NDC, S_{max}, E_{max}, F_{rate}, B, DPF, SDR_{max}). The model aims to represent:

- Saturation excess surface runoff;
- Threshold-based infiltration;
- Threshold-based baseflow;
- Deep percolation and water rise to meet evaporation demand.

S2.20.1 File names

Model: m_20_gsfb_8p_3s
Parameter ranges: m_20_gsfb_8p_3s_parameter_ ranges

S2.20.2 Model equations

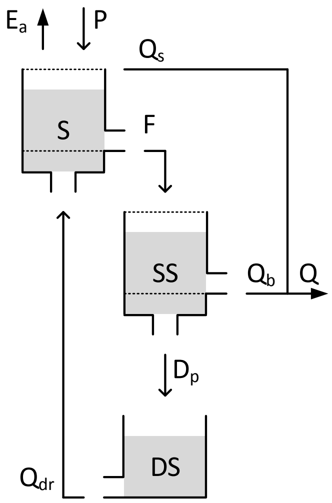

Figure S21: Structure of the GSFB model

$$\frac{dS}{dt} = P + Q_{dr} - E_a - Q_s - F \tag{189}$$

$$Q_{dr} = \begin{cases} C * DS * \left(1 - \frac{S}{NDC*S_{max}}\right), & \text{if } S \leq NDC * S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{190}$$

$$E_a = \begin{cases} E_p, & \text{if } S > NDC * S_{max} \\ min\left(E_p, E_{max} \frac{S}{NDC*S_{max}}\right), & \text{otherwise} \end{cases} \tag{191}$$

$$Q_s = \begin{cases} P, & \text{if } S = S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{192}$$

$$F = \begin{cases} F_{rate}, & \text{if } S > NDC * S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{193}$$

Where S [mm] is the current storage in the upper zone, refilled by precipitation P [mm/d] and recharge from deep groundwater Q_{dr} [mm/d]. The store is drained by evaporation E_a [mm/d], surface runoff Q_s [mm/d] and infiltration F [mm/d]. E_a occurs at the potential rate E_p [mm/d] if the store is above a threshold capacity given as the fraction NDC [-] of maximum storage S_{max} [mm]. Evaporation occurs at a reduced rate scaled by maximum evaporation rate E_{max} $[mm/d]$ if the store is below this threshold. Q_s occurs only if the store is at maximum capacity S_{max}. F occurs at a constant rate F_{rate} if the store is above threshold $NDC * S_{max}$. Recharge from deep percolation only occurs if the store is below threshold capacity $NDC * S_{max}$ and uses time parameter C $[d^{-1}]$ and current deep storage DS [mm].

$$\frac{dSS}{dt} = F - Q_b - D_p \tag{194}$$

$$Q_b = \begin{cases} B * DPF * (SS - SDR_{max}), & \text{if } SS > SDR_{max} \\ 0, & \text{otherwise} \end{cases} \tag{195}$$

$$D_p = (1 - B) * DPF * SS \tag{196}$$

Where SS [mm] is the current storage in the subsurface store, refilled by infiltration F and drained by baseflow Q_b $[mm/d]$ and deep percolation D_p $[mm/d]$. Outflow from this store is given as a function of storage DS and time coefficient DPF $[d^{-1}]$. A fraction $1 - B$ [-] of this outflow is deep percolation D_p. The remaining fraction B [-] is baseflow Q_b, provided the store is above threshold SDR_{max} [mm].

$$\frac{dDS}{dt} = D_p - Q_{dr} \tag{197}$$

$$\tag{198}$$

Where DS [mm] is the current storage in the deep store, refilled by a deep percolation D_p and drained by recharge to the upper store Q_{dr}. Total flow:

$$Q_t = Q_s + Q_b \tag{199}$$

S2.21 Flex-B (model ID: 21)

The Flex-B model (fig. S22) is the basis of a model development study (Fenicia et al., 2008). It has 3 stores and 9 parameters (UR_{max}, β, D, $Perc_{max}$, L_p,$N_{lag,f}$, $N_{lag,s}$, K_f, K_s). The model aims to represent:

- Infiltration and saturation excess flow based on a distribution of different soil depths;
- A split between fast saturation excess flow and preferential recharge to a slow store;
- Percolation from the unsaturated zone to a slow runoff store.

S2.21.1 File names

Model: m_21_flexb_9p_3s
Parameter ranges: m_21_flexb_9p_3s_parameter_ ranges

S2.21.2 Model equations

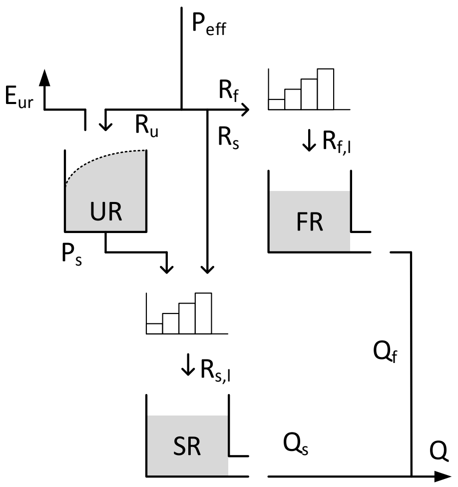

Figure S22: Structure of the Flex-B model

$$\frac{dUR}{dt} = R_u - E_{ur} - R_p \tag{200}$$

$$R_U = (1 - C_r) * P_{eff} \tag{201}$$

$$C_r = \left[1 + exp\left(\frac{-UR/UR_{max} + 1/2}{\beta}\right)\right]^{-1} \tag{202}$$

$$E_{ur} = E_p * min\left(1, \frac{UR}{UR_{max}}\frac{1}{L_p}\right) \tag{203}$$

$$P_s = Perc_{max} * \frac{UR}{UR_{max}} \tag{204}$$

Where UR is the current storage in the unsaturated zone [mm]. R_u $[mm/d]$ is the inflow into UR based on its current storage compared to maximum storage UR_{max} [mm] and a shape distribution parameter β [-].

E_{ur} the evaporation $[mm/d]$ from UR which follows a linear relation between current and maximum storage until a threshold L_p [-] is exceeded. P_s is the percolation from UR to the slow reservoir SR $[mm/d]$, based on a maximum percolation rate $Perc_{max}$ [mm], relative to the fraction of current storage and maximum storage. P_{eff} is routed towards the unsaturated zone based on Cr, with the remainder being divided into preferential recharge R_s $[mm/d]$ and fast runoff R_f $[mm/d]$:

$$R_s = (P_{eff} - R_u) * D \quad (205)$$
$$R_f = (P_{eff} - R_u) * (1 - D) \quad (206)$$

Where R_s and R_f are the flows $[mm/d]$ to the slow and fast runoff reservoir respectively, based on runoff partitioning coefficient D [-]. Both are lagged by linearly increasing triangular transformation functions with parameters $N_{lag,s}$ [d] and $N_{lag,f}$ [d] respectively. Percolation R_p is added to R_s before the transformation to $R_{s,l}$ occurs.

$$\frac{dFR}{dt} = R_{f,l} - Q_f \quad (207)$$
$$Q_f = K_f * FR \quad (208)$$

Where FR is the current storage [mm] in the fast flow reservoir. Outflow Q_f $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_f $[d^{-1}]$.

$$\frac{dSR}{dt} = R_{s,l} - Q_s \quad (209)$$
$$Q_s = K_s * SR \quad (210)$$

Where SR is the current storage [mm] in the slow flow reservoir. Outflow Q_s $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_s $[d^{-1}]$. Total outflow Q $[mm/d]$:

$$Q = Q_f + Q_s \quad (211)$$

S2.22 Variable Infiltration Capacity (VIC) model (model ID: 22)

The VIC model (fig. S23) is originally developed for use with General Circulation Models and uses latent and sensible heat fluxes to determine the rainfall-runoff relationship (Liang et al., 1994). For consistency with other models in this framework, we use a conceptualized version based in part of the VIC implementation in Clark et al. (2008). In addition, the original Leaf-Area-Index-based interception capacity is replaced with a sinusoidal curve-based approximation of interception capacity. The model has 3 stores and 10 parameters ($\bar{I}$, I_δ, I_s, $S_{sm,max}$, b, k_1, c_1, $S_{gw,max}$, k_2, c_2). The model aims to represent:

- Time-varying interception by vegetation;
- Variable infiltration and saturation excess flow;
- Interflow and baseflow from a deeper groundwater layer.

S2.22.1 File names

Model:	m_22_vic_10p_3s
Parameter ranges:	m_22_vic_10p_3s_parameter_ ranges

S2.22.2 Model equations

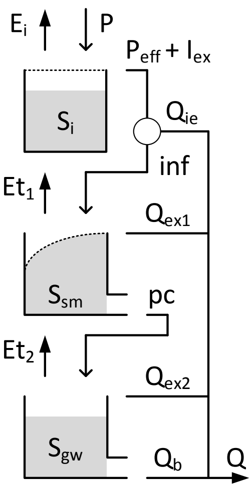

Figure S23: Structure of the VIC model

$$\frac{dS_i}{dt} = P - E_i - P_{eff} - I_{ex} \tag{212}$$

$$E_i = \frac{S_i}{I_{max}} * E_p \tag{213}$$

$$I_{max} = \bar{I}\left(1 + I_\delta * sin\left(2\pi(t + I_s)\right)\right) \tag{214}$$

$$P_eff = \begin{cases} P, & \text{if } S_i = I_{max} \\ 0, & \text{otherwise} \end{cases} \tag{215}$$

$$I_{ex} = max\left(S_i - I_{max}\right) \tag{216}$$

Where S_i [mm] is the current interception storage, refilled by precipitation P $[mm/d]$ and drained by evaporation E_i $[mm/d]$ and interception excess flows $P_{eff}[mm/d]$ and I_{ex} $[mm/d]$. E_i decreases linearly with storage, based on maximum storage I_{max} [mm]. I_{max} is determined using the mean interception $\bar{I}$ [mm], fractional seasonal interception change I_δ [-] and time shift I_s [-]. It is implicitly assumed that 1 sinusoidal period corresponds with a growing season of 1 year. P_{eff} is effective rainfall when the store is at maximum capacity. I_{ex} is an auxiliary flux used when a change in storage size result in current storage S_i exceeding I_{max}.

$$\frac{dS_{sm}}{dt} = inf - Et_1 - Q_{ex1} - pc \tag{217}$$

$$inf = (P_{eff} + I_{ex}) - Q_{ie} \tag{218}$$

$$Q_{ie} = (P_{eff} + I_{ex}) * \left(1 - \left(1 - \frac{S_{sm}}{S_{sm,max}}\right)^b\right) \tag{219}$$

$$E_{t1} = \frac{S_{sm}}{S_{sm,max}} * (E_p - E_i) \tag{220}$$

$$Q_{ex1} = \begin{cases} inf, & \text{if } S_{sm} = S_{sm,max} \\ 0, & \text{otherwise} \end{cases} \tag{221}$$

$$pc = k_1 * \left(\frac{S_{sm}}{S_{sm,max}}\right)^{c_1} \tag{222}$$

Where S_{sm} [mm] is the current soil moisture storage, refilled by infiltration inf $[mm/d]$, and drained by evapotranspiration Et_1 $[mm/d]$, storage excess Q_{ex1} $[mm/d]$ and percolation pc $[mm/d]$. inf relies on the value of infiltration excess Q_{ie}, which is calculated using the maximum soil moisture storage $S_{sm,max}$ [mm] and shape parameter b [-]. Et_1 scales linearly with current storage. Q_{ex1} equals inf when the store is at maximum capacity. pc has a potentially non-linear relationship with current storage through time parameter k_1 $[d^{-1}]$ and shape parameter c_1.

$$\frac{dS_{gw}}{dt} = pc - Et_2 - Q_{ex2} - Q_b \tag{223}$$

$$Et_2 = \frac{S_{gw}}{S_{gw,max}} * (E_p - E_i - Et_1) \tag{224}$$

$$Q_{ex2} = \begin{cases} pc, & \text{if } S_{gw} = S_{gw,max} \\ 0, & \text{otherwise} \end{cases} \tag{225}$$

$$Q_b = k_2 * \left(\frac{S_{gw}}{S_{gw,max}}\right)^{c_2} \tag{226}$$

Where S_{gw} [mm] is the current groundwater storage, refilled through percolation pc $[mm/d]$ and drained by evapotranspiration Et_2 $[mm/d]$, excess flow Q_{ex2} $[mm/d]$ and baseflow Q_b $[mm/d]$. Et_2 is scaled linearly with current storage based on maximum storage $S_{gw,max}$ [mm]. Q_{ex2} equals pc when the store is at maximum capacity. Q_b has a potentially non-linear relationship with current storage through time parameter k_2 and shape parameter c_2. Total outflow:

$$Q_t = Q_{ie} + Q_{ex1} + Q_{ex2} + Q_b \tag{227}$$

S2.23 Large-scale catchment water and salt balance model element (model ID: 23)

The large-scale catchment water and salt balance model (LASCAM) (fig. S24) is part of a study that investigates soil water and salt concentration before and after forest clearing (Sivapalan et al., 1996). It is a semi-distributed model made up of individual elements, such as described below. The model presented here simulates the water balance only (salt is ignored). It has 3 stores and 24 parameters (α_f, β_f, B_{max}, F_{max}, α_c, β_c, A_{min}, A_{max}, α_{ss}, β_{ss}, c, α_g, β_g, γ_f, δ_f, t_d, α_b, β_b, γ_a, δ_a, α_a, β_a, γ_b, δ_b). The model aims to represent:

- Stylized interception;
- Saturation and infiltration excess surface runoff;
- An inner layout representing near-stream saturated storage, deep saturated storage and medium-depth unsaturated storage;
- Subsurface saturation and infiltration excess flow to the near-stream store;
- Percolation to and capillary rise from groundwater.

S2.23.1 File names

Model: m_23_lascam_24p_3s
Parameter ranges: m_23_lascam_24p_3s_parameter_ ranges

S2.23.2 Model equations

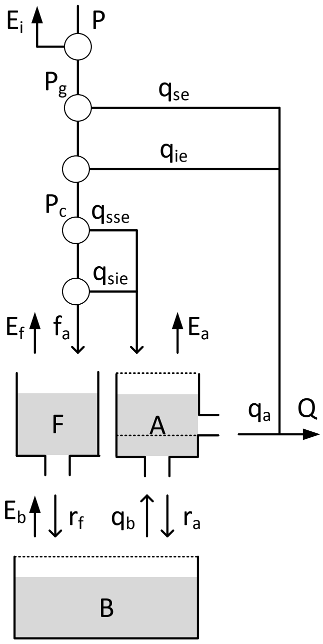

Figure S24: Structure of the LASCAM model

$$\frac{dF}{dt} = f_a - E_f - r_f \tag{228}$$

$$f_a = min\left(P_c * max\left(1, \frac{1-\phi_{ss}}{1-\phi_c}\right), f^*_{ss}\right) \tag{229}$$

$$f^*_{ss} = \alpha_f \left(1 - \frac{B}{B_{max}}\right)\left(\frac{F}{F_{max}}\right)^{-\beta_f} \tag{230}$$

$$\phi_c = \begin{cases} \alpha_c \left(\frac{A-A_{min}}{A_{max}-A_{min}}\right)^{\beta_c}, & \text{if } A > A_{min} \\ 0, & \text{otherwise} \end{cases} \tag{231}$$

$$\phi_{ss} = \begin{cases} \alpha_{ss} \left(\frac{A-A_{min}}{A_{max}-A_{min}}\right)^{\beta_{ss}}, & \text{if } A > A_{min} \\ 0, & \text{otherwise} \end{cases} \tag{232}$$

$$P_c = min\left(P_g - q_{se}, f^*_s\right) \tag{233}$$

$$f^*_s = c \tag{234}$$

$$q_{se} = \phi_c * P_g \tag{235}$$

$$P_g = max\left(\alpha_g + \beta_g * P, 0\right) \tag{236}$$

$$E_f = \gamma_f * E_p \left(\frac{F}{F_{max}}\right)^{\delta_f} \tag{237}$$

$$r_f = t_d * F \tag{238}$$

Where F [mm] is the current storage in the unsaturated infiltration store, which controls the amount of subsurface runoff generated on the boundary of a more permeable top layer (store A) with a less permeable bottom layer (store F). F is refilled by actual infiltration f_a $[mm/d]$, and drained by recharge r_f $[mm/d]$ and evaporation E_b $[mm/d]$. f_a depends on the actual infiltration rate P_c $[mm/d]$, the fraction saturated catchment area ϕ_{ss} [-], the fraction variable area contributing to overland flow ϕ_c [-] and a catchment-scale infiltration capacity f^*_{ss} $[mm/d]$. f^*_{ss} depends on a scaling parameter α_f $[mm/d]$, the relative storage in groundwater B/B_{max}, the relative infiltration volume in the catchment F/F_{max} and non-linearity parameter β_t [-]. B_{max} [mm] and F_{max} [mm] are storage scaling parameters [-]. ϕ_c uses the minimum contributing area A_{min} [mm], maximum contributing area A_{max} [mm] and shape parameters α_c [-] and β_c [-] to control the shape of this distribution. ϕ_{ss} takes a similar shape as ϕ_c, using parameters α_{ss} [-] and β_{ss} [-]. P_c is the lesser of throughfall rate P_g $[mm/d]$ minus saturation excess q_{se} $[mm/d]$, and the catchment infiltration capacity f^*_s $[mm/d]$. f^*_s is assumed to have a constant rate c $[mm/d]$. q_{se} is determined as that part of throughfall P_g that falls on the variable contributing catchment area given by ϕ_c. P_g is determined as a fixed interception rate α_g $[mm/d]$ and a fractional interception β_g [-]. Evaporation E_f uses the potential rate E_p $[mm/d]$ scaled by the relative storage in F and two shape parameters γ_f [-] and δ_f [-]. Recharge r_f $[mm/d]$ has a linear relation with storage through time parameter t_d $[d^{-1}]$.

$$\frac{dA}{dt} = q_{sse} + q_{sie} + q_b - E_a - q_a - r_a \tag{239}$$

$$q_{sse} = \frac{\phi_{ss} - \phi_c}{1 - \phi_c} P_c \tag{240}$$

$$q_{sie} = max\left(P_c * \frac{1 - \phi_{ss}}{1 - \phi_c} - f^*_{ss}, 0\right) \tag{241}$$

$$q_b = \beta_b \left(exp\left(\alpha_b \frac{B}{B_{max}}\right) - 1\right) \tag{242}$$

$$E_a = \phi_c * E_p + \gamma_a * E_p \left(\frac{A}{A_{max}}\right)^{\delta_a} \tag{243}$$

$$q_a = \begin{cases} \alpha_a \left(\frac{A - A_{min}}{A_{max} - A_{min}}\right)^{\beta_a}, & \text{if } A > A_{min} \\ 0, & \text{otherwise} \end{cases} \tag{244}$$

$$r_a = \phi_{ss} * f^*_{ss} \tag{245}$$

Where A [mm] is the current storage in the more permeable upper zone (above less permeable lower zone F), refilled by sub-surface saturation excess q_{sse} $[mm/d]$, sub-surface infiltration excess q_{sie} $[mm/d]$ and discharge from groundwater q_b $[mm/d]$. The store is drained by evaporation E_a, subsurface stormflow q_a $[mm/d]$ and recharge r_a $[mm/d]$. Flow from store B, q_b, decreases exponentially as the store dries out, controlled by parameters β_b and α_b. Evaporation E_a occurs at the potential rate E_p from the variable saturated area ϕ_c and additionally at a rate scaled by the relative storage in A and two shape parameters γ_a [-] and $delta_a$ [-]/ Recharge r_a is a function of the saturated subsurface area ϕ_{ss} and the subsurface infiltration rate f^*_{ss}.

$$\frac{dB}{dt} = r_f + r_a - E_b - q_b \tag{246}$$

$$E_b = \gamma_b * E_p \left(\frac{B}{B_{max}}\right)^{\delta_b} \tag{247}$$

Where B [mm] is the current storage in the deep layers, refilled by recharge from stores A (r_a) and F (r_f), and drained by evaporation E_b and groundwater discharge q_b. E_b uses the potential rate E_p scaled by the relative storage in B and two shape parameters γ_b [-] and δ_b [-]. Total flow:

$$Q_t = q_{se} + q_{ie} + q_a \tag{248}$$

$$q_{ie} = P_g - q_{se} - P_c \tag{249}$$

Where q_{ie} $[mm/d]$ is infiltration excess on the surface.

S2.24 MOPEX-1 (model ID: 24)

The MOPEX-1 model (fig. S25) is part of a model improvement study that investigates the relationship between dominant processes and model structures for 197 catchments in the MOPEX database (Ye et al., 2012). It has 4 stores and 5 parameters (S_{b1}, t_w, t_u, S_e, t_c). The model aims to represent:

- Saturation excess flow;
- Infiltration to deeper soil layers;
- A split between fast and slow runoff.

S2.24.1 File names

Model: m_24_mopex1_5p_4s
Parameter ranges: m_24_mopex1_5p_4s_parameter_ ranges

S2.24.2 Model equations

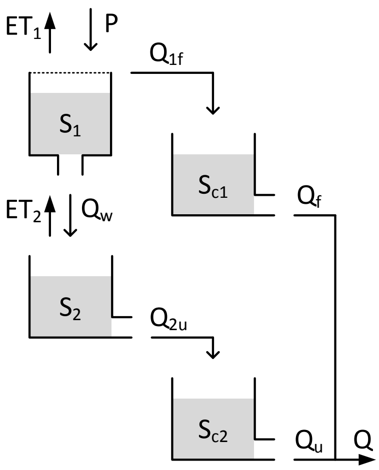

Figure S25: Structure of the MOPEX-1 model

$$\frac{dS_1}{dt} = P - ET_1 - Q_{1f} - Q_w \tag{250}$$

$$ET_1 = \frac{S_1}{S_{b1}} * Ep \tag{251}$$

$$Q_{1f} = \begin{cases} P, & \text{if } S_1 \geq S_{b1} \\ 0, & \text{otherwise} \end{cases} \tag{252}$$

$$Q_w = t_w * S_1 \tag{253}$$

Where S_1 [mm] is the current storage in soil moisture and P precipitation $[mm/d]$. Evaporation ET_1 $[mm/d]$ depends linearly on current soil moisture, maximum soil moisture S_{b1} [mm] and potential evapotransporation E_p [mm/d]. Saturation excess flow Q_{1f} $[mm/d]$ occurs when the soil moisture bucket exceeds its maximum capacity. Infiltration to deeper groundwater Q_w $[mm/d]$ depends on current soil moisture and time parameter t_w $[d^{-1}]$.

$$\frac{dS_2}{dt} = Q_w - ET_2 - Q_{2u} \tag{254}$$

$$ET_2 = \frac{S_2}{S_e} * Ep \tag{255}$$

$$Q_{2u} = t_u * S_2 \tag{256}$$

Where S_2 [mm] is the current groundwater storage, refilled by infiltration from S_1. Evaporation ET_2 $[mm/d]$ depends linearly on current groundwater and root zone storage capacity S_e [mm]. Leakage to the slow runoff store Q_{2u} $[mm/d]$ depends on current groundwater level and time parameter t_u $[d^{-1}]$.

$$\frac{dS_{c1}}{dt} = Q_{1f} - Q_f \tag{257}$$

$$Q_f = t_c * S_{c1} \tag{258}$$

Where S_{c1} [mm] is current storage in the fast flow routing reservoir, refilled by Q_{1f}. Routed flow Q_f depends on the mean residence time parameter t_c $[d^{-1}]$.

$$\frac{dS_{c2}}{dt} = Q_{2u} - Q_u \tag{259}$$

$$Q_u = t_c * S_{c2} \tag{260}$$

Where S_{c2} [mm] is current storage in the slow flow routing reservoir, refilled by Q_{2u}. Routed flow Q_u depends on the mean residence time parameter t_c $[d^{-1}]$. Total simulated flow Q_t $[mm/d]$:

$$Q_t = Q_f + Q_u \tag{261}$$

S2.25 Thames Catchment Model (model ID: 25)

The Thames Catchment Model (TCM) model (fig. S26) is originally intended to be used in zones with similar surface characteristics, rather than catchments as a whole (Moore and Bell, 2001). It has 4 stores and 6 parameters (ϕ, rc, γ, k_1, c_a, k_2). The model aims to represent:

- Effective rainfall before infiltration;
- Preferential recharge;
- Catchment drying through prolonged soil moisture depletion;
- Groundwater abstraction;
- Non-linear groundwater flow.

S2.25.1 File names

Model: m_25_tcm_6p_4s
Parameter ranges: m_25_tcm_6p_4s_parameter_ ranges

S2.25.2 Model equations

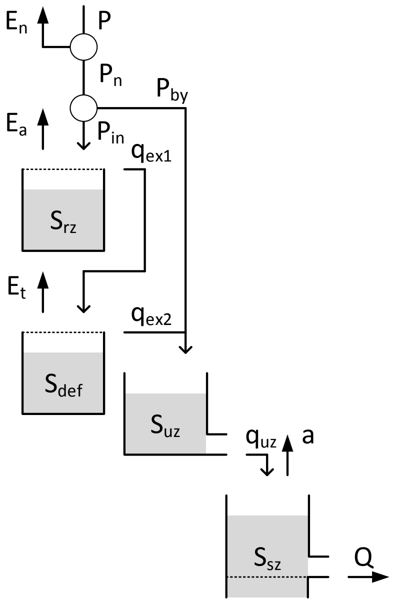

Figure S26: Structure of the TCM model

$$\frac{dS_{Rz}}{dt} = P_{in} - E_a - q_{ex1} \tag{262}$$

$$P_{in} = (1-\phi) * P_n \tag{263}$$

$$P_n = max(P - E_p, 0) \tag{264}$$

$$E_a = \begin{cases} E_p, & \text{if } S_{rz} > 0 \\ 0, & \text{otherwise} \end{cases} \tag{265}$$

$$q_{ex1} = \begin{cases} P_{in}, & \text{if } S_{rz} > rc \\ 0, & \text{otherwise} \end{cases} \tag{266}$$

Where S_{rz} [mm] is the current storage in the root zone, refilled by infiltrated precipitation P_{in} $[mm/d]$, and drained by evaporation E_a $[mm/d]$ and storage excess flow q_{ex1} $[mm/d]$. P_{in} is the fraction $(1-\phi)$ [-] of net precipitation P_n $[mm/d]$ that is not preferential recharge. P_n is the difference between precipitation P [mm/d] and potential evapotranspiration E_p [mm/d] per time step. E_a occurs at the net potential rate whenever possible. q_{ex1} occurs only when the store is at maximum capacity rc [mm].

$$\frac{dS_{def}}{dt} = E_t + q_{ex2} - q_{ex1} \tag{267}$$

$$E_t = \begin{cases} \gamma * E_p, & \text{if } S_{rz} = 0 \\ 0, & \text{otherwise} \end{cases} \tag{268}$$

$$q_{ex2} = \begin{cases} q_{ex1}, & \text{if } S_{def} = 0 \\ 0, & \text{otherwise} \end{cases} \tag{269}$$

Where S_{def} [mm] is the current storage in the soil moisture *deficit* store. The deficit is increased by evaporation E_t $[mm/d]$ and percolation q_{ex2} $[mm/d]$. The deficit is decreased by overflow from the upper store q_{ex1}. E_t only occurs when the upper zone is empty and at a fraction γ [-] of E_p. q_{ex2} only occurs when the deficit is zero.

$$\frac{dS_{uz}}{dt} = P_{by} + q_{ex2} - q_{uz} \tag{270}$$

$$P_{by} = \phi * P_n \tag{271}$$

$$q_{uz} = k_1 * S_{uz} \tag{272}$$

Where S_{uz} is the current storage in the unsaturated zone, refilled by preferential recharge P_{by} $[mm/d]$ and percolation q_{ex2} $[mm/d]$, and drained by groundwater flow q_{uz} $[mm/d]$. P_{by} is a fraction ϕ [-] of P_n. q_{uz} has a linear relation with storage through time parameter k_1 $[d^{-1}]$.

$$\frac{dS_{sz}}{dt} = q_{uz} - a - Q \tag{273}$$

$$a = c_a \tag{274}$$

$$Q = \begin{cases} k_2 * S_{sz}^2, & \text{if } S_{sz} > 0 \\ 0, & \text{otherwise} \end{cases} \tag{275}$$

Where S_{sz} [mm] is the current storage in the saturated zone, refilled by groundwater flow q_{uz} $[mm/d]$ and drained by abstractions a $[mm/d]$ and outflow Q $[mm/d]$. a occurs at a constant rate c_a $[mm/d]$. Abstractions can draw down the aquifer below the runoff generating threshold. Q has a quadratic relation with storage through parameter k_2 $[mm^{-1}d^{-1}]$.

S2.26 Flex-I (model ID: 26)

The Flex-I model (fig. S27) is the part of a model development exercise (Fenicia et al., 2008). It has 4 stores and 10 parameters (I_{max}, UR_{max}, β, D, $Perc_{max}$, L_p,$N_{lag,f}$, $N_{lag,s}$, K_f, K_s). The model aims to represent:

- Interception by vegetation;
- Infiltration and saturation excess flow based on a distribution of different soil depths;
- A split between fast saturation excess flow and preferential recharge to a slow store;
- Percolation from the unsaturated zone to a slow runoff store.

S2.26.1 File names

Model: m_26_flexi_10p_4s
Parameter ranges: m_26_flexi_10p_4s_parameter_ ranges

S2.26.2 Model equations

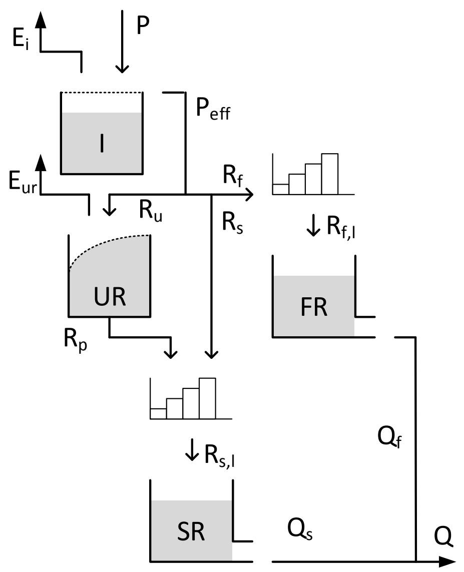

Figure S27: Structure of the Flex-I model

$$\frac{dI}{dt} = P - E_i - P_{eff} \tag{276}$$

$$E_i = \begin{cases} Ep & \text{, if } I > 0 \\ 0 & \text{, otherwise} \end{cases} \tag{277}$$

$$P_eff = \begin{cases} P & \text{, if } I \geq I_{max} \\ 0 & \text{, otherwise} \end{cases} \tag{278}$$

Where I is the current interception storage [mm], P $[mm/d]$ incoming precipitation, E_i $[mm/d]$ evaporation from the interception store and P_{eff} $[mm/d]$ interception excess routed to soil moisture. Evaporation occurs at the potential rate E_p [mm/d] whenever possible. Interception excess occurs when the interception store exceeds its maximum capacity I_{max} [mm].

$$\frac{dUR}{dt} = R_u - E_{ur} - R_p \tag{279}$$
$$R_U = (1 - C_r) * P_{eff} \tag{280}$$
$$C_r = \Big[1 + exp\Big(\frac{-UR/UR_{max} + 1/2}{\beta}\Big)\Big]^{-1} \tag{281}$$
$$E_{ur} = E_p * min\Big(1, \frac{UR}{UR_{max}}\frac{1}{L_p}\Big) \tag{282}$$
$$P_s = Perc_{max} * \frac{-UR}{UR_{max}} \tag{283}$$

Where UR is the current storage in the unsaturated zone [mm]. R_u $[mm/d]$ is the inflow into UR based on its current storage compared to maximum storage UR_{max} [mm] and a shape distribution parameter β [-]. E_{ur} the evaporation $[mm/d]$ from UR which follows a linear relation between current and maximum storage until a threshold L_p [-] is exceeded. P_s is the percolation from UR to the slow reservoir SR $[mm/d]$, based on a maximum percolation rate $Perc_{max}$ [mm], relative to the fraction of current storage and maximum storage. P_{eff} is routed towards the unsaturated zone based on Cr, with the remainder being divided into preferential recharge R_s $[mm/d]$ and fast runoff R_f $[mm/d]$:

$$R_s = (P_{eff} - R_u) * D \tag{284}$$
$$R_f = (P_{eff} - R_u) * (1 - D) \tag{285}$$

Where R_s and R_f are the flows $[mm/d]$ to the slow and fast runoff reservoir respectively, based on runoff partitioning coefficient D [-]. Both are lagged by linearly increasing triangular transformation functions with parameters $N_{lag,s}$ [d] and $N_{lag,f}$ [d] respectively, that give the number of days over which R_s and R_f need to be transformed. Percolation R_p is added to R_s before the transformation to $R_{s,l}$ occurs.

$$\frac{dFR}{dt} = R_{f,l} - Q_f \tag{286}$$
$$Q_f = K_f * FR \tag{287}$$

Where FR is the current storage [mm] in the fast flow reservoir. Outflow Q_f $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_f $[d^{-1}]$.

$$\frac{dSR}{dt} = R_{s,l} - Q_s \tag{288}$$
$$Q_s = K_s * SR \tag{289}$$

Where SR is the current storage [mm] in the slow flow reservoir. Outflow Q_s $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_s $[d^{-1}]$. Total outflow Q $[mm/d]$:

$$Q = Q_f + Q_s \tag{290}$$

S2.27 Tank model (model ID: 27)

The Tank Model (fig. S28) is originally developed for use constantly saturated soils in Japan (Sugawara, 1995). It has 4 stores and 12 parameters (A_0, A_1, A_2, t_1, t_2, B_0, B_1, t_3, C_0, C_1, t_4, D_1). The model aims to represent:

- Runoff on increasing time scales with depth.

S2.27.1 File names

Model: m_27_tank_12p_4s
Parameter ranges: m_27_tank_12p_4s_parameter_ ranges

S2.27.2 Model equations

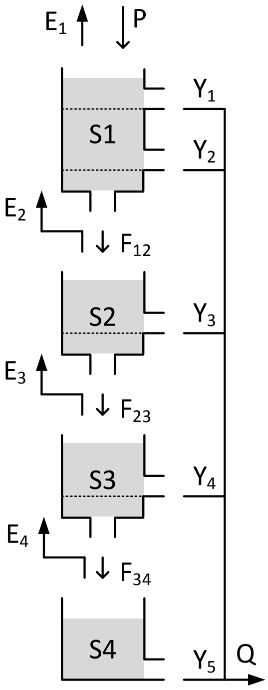

Figure S28: Structure of the Tank Model

$$\frac{dS_1}{dt} = P - E_1 - F_{12} - Y_2 - Y_1 \tag{291}$$

$$E_1 = \begin{cases} Ep, & \text{if } S_1 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{292}$$

$$F_{12} = A_0 * S_1 \tag{293}$$

$$Y_2 = \begin{cases} A_2 * (S_1 - t_2), & \text{if } S_1 > t_2 \\ 0, & \text{otherwise} \end{cases} \tag{294}$$

$$Y_1 = \begin{cases} A_1 * (S_1 - t_1), & \text{if } S_1 > t_1 \\ 0, & \text{otherwise} \end{cases} \tag{295}$$

Where S_1 [mm] is the current storage in the upper zone, refilled by precipitation P $[mm/d]$ and drained by evaporation E_1 $[mm/d]$, drainage F_{12} $[mm/d]$ and surface runoff Y_1 $[mm/d]$ and Y_2 $[mm/d]$. E_1 occurs at the potential rate E_p $[mm/d]$ if water is available. Drainage to the intermediate layer has a linear relationship with storage through time scale parameter A_0 $[d^{-1}]$. Surface runoff Y_2 and Y_1 occur when S_1 is above thresholds t_2 [mm] and t_1 [mm] respectively. Both are linear relationships through time parameters A_2 $[d^{-1}]$ and A_1 $[d^{-1}]$ respectively.

$$\frac{dS_2}{dt} = F_{12} - E_2 - F_{23} - Y_3 \tag{296}$$

$$E_2 = \begin{cases} Ep, & \text{if } S_1 = 0 \ \& \ S_2 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{297}$$

$$F_{23} = B_0 * S_2 \tag{298}$$

$$Y_3 = \begin{cases} B_1 * (S_2 - t_3), & \text{if } S_2 > t_3 \\ 0, & \text{otherwise} \end{cases} \tag{299}$$

Where S_2 [mm] is the current storage in the intermediate zone, refilled by drainage F_{12} from the upper zone and drained by evaporation E_2 $[mm/d]$, drainage F_{23} $[mm/d]$ and intermediate discharge Y_3 $[mm/d]$. E_2 occurs at the potential rate E_p if water is available and the upper zone is empty. Drainage to the third layer F_{23} has a linear relationship with storage through time scale parameter B_0 $[d^{-1}]$. Intermediate runoff Y_3 occurs when S_2 is above threshold t_3 [mm] and has a linear relationship with storage through time scale parameter B_1 $[d^{-1}]$.

$$\frac{dS_3}{dt} = F_{23} - E_3 - F_{34} - Y_4 \tag{300}$$

$$E_3 = \begin{cases} Ep, & \text{if } S_1 = 0 \ \& \ S_2 = 0 \ \& \ S_3 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{301}$$

$$F_{34} = C_0 * S_3 \tag{302}$$

$$Y_4 = \begin{cases} C_1 * (S_3 - t_4), & \text{if } S_3 > t_4 \\ 0, & \text{otherwise} \end{cases} \tag{303}$$

Where S_3 [mm] is the current storage in the sub-base zone, refilled by drainage F_{23} from the intermediate zone and drained by evaporation E_3 $[mm/d]$, drainage F_{34} $[mm/d]$ and sub-base discharge Y_4 $[mm/d]$. E_3 occurs at the potential rate E_p if water is available and the upper zones are empty. Drainage to the fourth layer F_{34} has a linear relationship with storage through time scale parameter C_0 $[d^{-1}]$. Sub-base runoff Y_4 occurs when S_3 is above threshold t_4 [mm] and has a linear relationship with storage through time scale parameter C_1 $[d^{-1}]$.

$$\frac{dS_4}{dt} = F_{34} - E_4 - Y_5 \tag{304}$$

$$E_4 = \begin{cases} Ep, & \text{if } S_1 = 0 \ \& \ S_2 = 0 \ \& \ S_3 = 0 \ \& \ S_4 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{305}$$

$$Y_5 = D_1 * S_4 \tag{306}$$

Where S_4 [mm] is the current storage in the base layer, refilled by drainage F_{34} from the sub-base zone and drained by evaporation E_4 $[mm/d]$ and baseflow Y_5 $[mm/d]$.

E_4 occurs at the potential rate E_p if water is available and the upper zones are empty. Baseflow Y_5 has a linear relationship with storage through time scale parameter D_1 $[d^{-1}]$. Total runoff:

$$Q_t = Y_1 + Y_2 + Y_3 + Y_4 + Y_5 \tag{307}$$

S2.28 Xinanjiang model (model ID: 28)

The Xinanjiang model (fig. S29) is originally intended for use in humid or semi-humid regions in China (Zhao, 1992). The model uses a variable contributing area to simulate runoff. The version presented here uses a double parabolic curve to simulate tension water capacities within the catchment (Jayawardena and Zhou, 2000), instead of the original single parabolic curve. The model has 4 stores and 12 parameters (A_{im}, a, b, W_{max}, LM, c, S_{max}, Ex, k_I, k_G, c_I, c_G). The model aims to represent:

- Runoff from impervious areas;
- Variable distribution of tension water storage capacities in the catchment;
- Variable contributing area of free water storages;
- Direct surface runoff from the contributing free area;
- Delayed interflow and baseflow from the contributing free area.

S2.28.1 File names

Model: m_28_xinanjiang_12p_4s
Parameter ranges: m_28_xinanjiang_12p_4s_parameter_ ranges

S2.28.2 Model equations

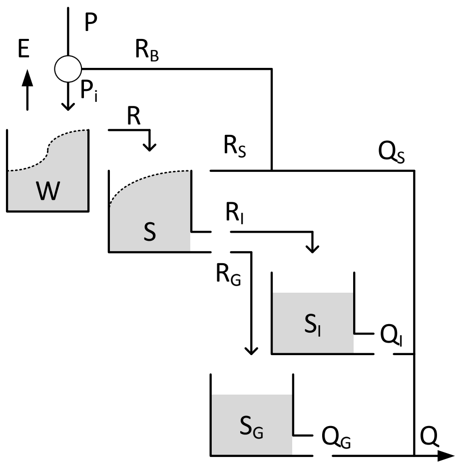

Figure S29: Structure of the Xinanjiang model

$$\frac{dW}{dt} = P_i - E - R \tag{308}$$

$$P_i = (1 - A_{im}) * P \tag{309}$$

$$R = \begin{cases} P_i * \left[(0.5-a)^{1-b}\left(\frac{W}{W_{max}}\right)^b\right], & \text{if } \frac{W}{W_{max}} \leq 0.5 - a \\ P_i * \left[1-(0.5+a)^{1-b}\left(1-\frac{W}{W_{max}}\right)^b\right], & \text{otherwise} \end{cases} \tag{310}$$

$$E = \begin{cases} E_p, & \text{if } W > LM \\ \frac{W}{LM}E_p, & \text{if } c*LM \geq W \leq LM \\ c*E_p, & \text{otherwise} \end{cases} \tag{311}$$

Where W [mm] is the current tension water storage, refilled by a infiltration P_i [mm/d] and drained by evaporation E [mm/d] and runoff R [mm/d]. P_i is the fraction of precipitation P [mm/d] that does not fall on impervious area A_{im} [-]. Runoff generation R uses a double parabolic curve to determine the fraction of catchment area that is at full tension storage and thus can contribute to runoff generation. This curve relies on shape parameters a [-] and b [-], and maximum tension water storage W_{max} [mm]. Evaporation rate E declines as tension water storage decreases. Evaporation occurs at the potential rate E_p $[mm/d]$ if storage W is above threshold LM [mm], and reduces linearly below that up to a second threshold $c * LM$ [-]*[mm]. Below this threshold evaporation occurs at a constant rate $c * E_p$.

$$\frac{dS}{dt} = R - R_S - R_I - R_G \tag{312}$$

$$R_S = R * \left(1 - \left(1 - \frac{S}{S_{max}}\right)^{Ex}\right) \tag{313}$$

$$R_I = k_I * S * \left(1 - \left(1 - \frac{S}{S_{max}}\right)^{Ex}\right) \tag{314}$$

$$R_G = k_G * S * \left(1 - \left(1 - \frac{S}{S_{max}}\right)^{Ex}\right) \tag{315}$$

Where S [mm] is the current storage of free water, refilled by runoff R from filled tension water areas, and drained by surface runoff R_S [mm/d], interflow R_I [mm/d] and baseflow R_G [mm/d]. All runoff components rely on a parabolic equation to simulate variable contributing areas of the catchment, dependent on maximum free water storage S_{max} [mm] and shape parameter Ex [-]. R_I also uses a time coeficient k_I $[d^{-1}]$. R_G uses a time coeficient k_G $[d^{-1}]$.

$$\frac{dS_I}{dt} = R_I - Q_I \tag{316}$$

$$Q_I = c_I * S_I \tag{317}$$

Where S_I [mm] is the current storage in the interflow routing reservoir, filled by interflow from free water R_I and drained by delayed interflow Q_I $[mm/d]$. Q_I uses a time coefficient c_I $[d^{-1}]$.

$$\frac{dS_G}{dt} = R_G - Q_G \tag{318}$$

$$Q_G = c_G * S_G \tag{319}$$

Where S_G [mm] is the current storage in the baseflow routing reservoir, filled by baseflow from free water R_G and drained by delayed baseflow Q_G $[mm/d]$. Q_G uses a time coefficient c_G $[d^{-1}]$. Total flow depends on four separate runoff components:

$$Q_t = Q_S + Q_I + Q_G \tag{320}$$

$$Q_S = R_S + R_B \tag{321}$$

$$R_B = A_{im} * P \tag{322}$$

Where R_B [mm/d] is direct rainoff generated by precipitation P [mm/d] on the fraction impervious area A_{im} [-].

S2.29 HyMOD (model ID: 29)

The HyMOD model (fig. S30) combines a PDM-like soil moisture routine (e.g. Moore (2007)) with a Nash cascade of three linear reservoirs that simulates fast flow and a single linear reservoir intended to simulate slow flow (Wagener et al., 2001; Boyle, 2001). Although the model was originally intended as a flexible structure where the user defines which processes to include, this study includes only a single version that is commonly used. It has 5 parameters (S_{max}, b, a, k_f and k_s) and 5 stores. The model aims to represent:

- Different soil depths throughout the catchment;
- Separation of flow into fast and slow flow.

S2.29.1 File names

Model:	m_29_hymod_5p_5s
Parameter ranges:	m_29_hymod_5p_5s_parameter_ ranges

S2.29.2 Model equations

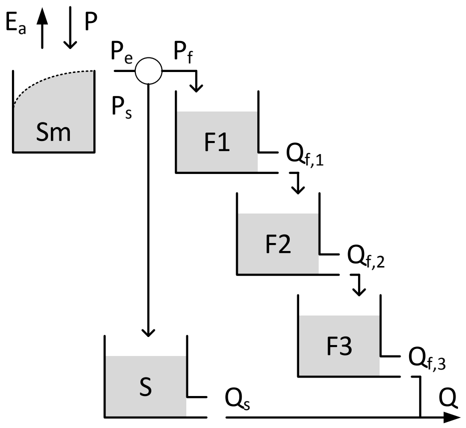

Figure S30: Structure of the HyMOD model

$$\frac{dSm}{dt} = P - E_a - P_e \quad (323)$$

$$E_a = \frac{Sm}{S_{max}} * E_p \quad (324)$$

$$P_e = \left(1 - \left(1 - \frac{S}{S_{max}}\right)^b\right) * P \quad (325)$$

Where Sm is the current storage in Sm [mm], S_{max} [mm] is the maximum storage in Sm, E_a and E_p the actual and potential evapotranspiration respectively [mm/d] and b is the soil depth distribution parameter [-]. P [mm/d] is the precipitation input.

$$\frac{dF_1}{dt} = P_f - Q_{f,1} \quad (326)$$

$$P_f = a * P_e \quad (327)$$

$$Q_{f,1} = k_f * S_{f,1} \quad (328)$$

Where F_1 is the current storage in store F_1 [mm], a the fraction of P_e that flows into the fast stores and k_f the runoff coefficient of the fast stores. Stores F_2 and F_3 take the outflow of the previous store as input ($Q_{f,1}$ and $Q_{f,2}$ respectively) and generate outflow analogous to the equations above.

$$\frac{dS}{dt} = P_s - Q_s \tag{329}$$

$$P_s = (1 - a) * P_e \tag{330}$$

$$Q_s = k_s * S \tag{331}$$

Where S is the current storage in store S [mm], $1 - a$ [-] the fraction of P_e that flows into the slow store and k_s the runoff coefficient of the slow store. Total outflow:

$$Q_t = Q_s + Q_{f,3} \tag{332}$$

S2.30 MOPEX-2 (model ID: 30)

The MOPEX-2 model (fig. S31) is part of a model improvement study that investigates the relationship between dominant processes and model structures for 197 catchments in the MOPEX database (Ye et al., 2012). It has 5 stores and 7 parameters (T_{crit}, ddf, S_{b1}, t_w, t_u, S_e, t_c). The model aims to represent:

- Snow accumulation and melt;
- Saturation excess flow;
- Infiltration to deeper soil layers;
- A split between fast and slow runoff.

S2.30.1 File names

Model: m_30_mopex2_7p_5s
Parameter ranges: m_30_mopex2_7p_5s_parameter_ ranges

S2.30.2 Model equations

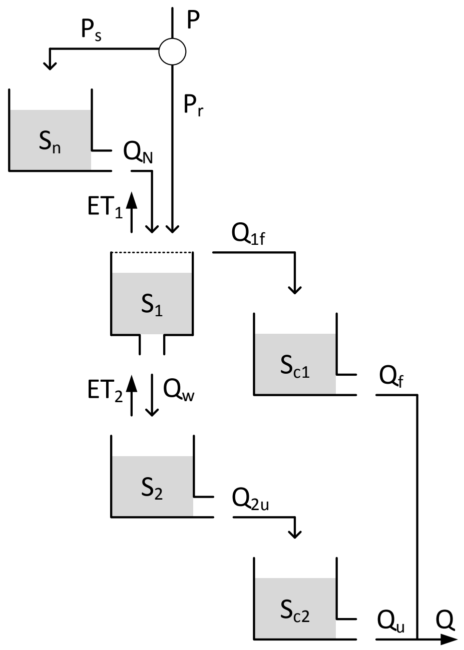

Figure S31: Structure of the MOPEX-2 model

$$\frac{dS_n}{dt} = P_s - Q_n \tag{333}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{334}$$

$$Q_n = \begin{cases} ddf * (T - T_{crit}), & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{335}$$

Where S_n [mm] is the current snow pack. Precipitation occurs as snowfall P_s $[mm/d]$ when current temperature T $[^oC]$ is below threshold T_{crit} $[^oC]$. Snowmelt Q_N $[mm/d]$ occurs when the temperature rises above the threshold temperature and relies in the degree-day factor dd $[mm/^oC/d]$.

$$\frac{dS_1}{dt} = P_r - ET_1 - Q_{1f} - Q_w \tag{336}$$

$$P_r = \begin{cases} P, & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{337}$$

$$ET_1 = \frac{S_1}{S_{b1}} * Ep \tag{338}$$

$$Q_{1f} = \begin{cases} P, & \text{if } S_1 \geq S_{b1} \\ 0, & \text{otherwise} \end{cases} \tag{339}$$

$$Q_w = t_w * S_1 \tag{340}$$

Where S_1 [mm] is the current storage in soil moisture and P_r precipitation as rain $[mm/d]$. Evaporation ET_1 $[mm/d]$ depends linearly on current soil moisture, maximum soil moisture S_{b1} [mm] and potential evapotranspiration E_p [mm/d]. Saturation excess flow Q_{1f} $[mm/d]$ occurs when the soil moisture bucket exceeds its maximum capacity. Infiltration to deeper groundwater Q_w $[mm/d]$ depends on current soil moisture and time parameter t_w $[d^{-1}]$.

$$\frac{dS_2}{dt} = Q_w - ET_2 - Q_{2u} \tag{341}$$

$$ET_2 = \frac{S_2}{S_e} * Ep \tag{342}$$

$$Q_{2u} = t_u * S_2 \tag{343}$$

Where S_2 [mm] is the current groundwater storage, refilled by infiltration from S_1. Evaporation ET_2 $[mm/d]$ depends linearly on current groundwater and root zone storage capacity S_e [mm]. Leakage to the slow runoff store Q_{2u} $[mm/d]$ depends on current groundwater level and time parameter t_u $[d^{-1}]$.

$$\frac{dS_{c1}}{dt} = Q_{1f} - Q_f \tag{344}$$

$$Q_f = t_c * S_{c1} \tag{345}$$

Where S_{c1} [mm] is current storage in the fast flow routing reservoir, refilled by Q_{1f}. Routed flow Q_f depends on the mean residence time parameter t_c $[d^{-1}]$.

$$\frac{dS_{c2}}{dt} = Q_{2u} - Q_u \tag{346}$$

$$Q_u = t_c * S_{c2} \tag{347}$$

Where S_{c2} [mm] is current storage in the slow flow routing reservoir, refilled by Q_{2u}. Routed flow Q_u depends on the mean residence time parameter t_c $[d^{-1}]$. Total simulated flow Q_t $[mm/d]$:

$$Q_t = Q_f + Q_u \tag{348}$$

S2.31 MOPEX-3 (model ID: 31)

The MOPEX-3 model (fig. S32) is part of a model improvement study that investigates the relationship between dominant processes and model structures for 197 catchments in the MOPEX database (Ye et al., 2012). It has 5 stores and 8 parameters (T_{crit}, ddf, S_{b1}, t_w, S_{b2}, t_u, S_e, t_c). The model aims to represent:

- Snow accumulation and melt;
- Saturation excess flow;
- Infiltration to deeper soil layers;
- Subsurface-influenced fast flow;
- A split between fast and slow runoff.

S2.31.1 File names

Model: m_31_mopex3_8p_5s
Parameter ranges: m_31_mopex3_8p_5s_parameter_ ranges

S2.31.2 Model equations

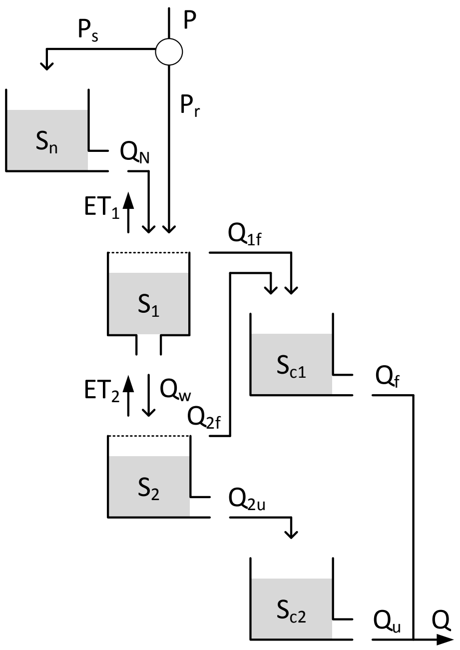

Figure S32: Structure of the MOPEX-3 model

$$\frac{dS_n}{dt} = P_s - Q_n \tag{349}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{350}$$

$$Q_n = \begin{cases} ddf * (T - T_{crit}), & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{351}$$

Where S_n [mm] is the current snow pack. Precipitation occurs as snowfall P_s $[mm/d]$ when current temperature T $[^oC]$ is below threshold T_{crit} $[^oC]$. Snowmelt Q_N $[mm/d]$ occurs when the temperature rises above the threshold temperature and relies in the degree-day factor dd $[mm/^oC/d]$.

$$\frac{dS_1}{dt} = P_r - ET_1 - Q_{1f} - Q_w \tag{352}$$

$$P_r = \begin{cases} P, & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{353}$$

$$ET_1 = \frac{S_1}{S_{b1}} * Ep \tag{354}$$

$$Q_{1f} = \begin{cases} P, & \text{if } S_1 \geq S_{b1} \\ 0, & \text{otherwise} \end{cases} \tag{355}$$

$$Q_w = t_w * S_1 \tag{356}$$

Where S_1 [mm] is the current storage in soil moisture and P_r precipitation as rain $[mm/d]$. Evaporation ET_1 $[mm/d]$ depends linearly on current soil moisture, maximum soil moisture S_{b1} [mm] and potential evapotranspiration E_p [mm/d]. Saturation excess flow Q_{1f} $[mm/d]$ occurs when the soil moisture bucket exceeds its maximum capacity. Infiltration to deeper groundwater Q_w $[mm/d]$ depends on current soil moisture and time parameter t_w $[d^{-1}]$.

$$\frac{dS_2}{dt} = Q_w - ET_2 - Q_{2u} - Q_{2f} \tag{357}$$

$$ET_2 = \frac{S_2}{S_e} * Ep \tag{358}$$

$$Q_{2u} = t_u * S_2 \tag{359}$$

$$Q_{2f} = \begin{cases} Q_w, & \text{if } S_2 \geq S_{b2} \\ 0, & \text{otherwise} \end{cases} \tag{360}$$

Where S_2 [mm] is the current groundwater storage, refilled by infiltration from S_1. Evaporation ET_2 $[mm/d]$ depends linearly on current groundwater and root zone storage capacity S_e [mm]. Leakage to the slow runoff store Q_{2u} $[mm/d]$ depends on current groundwater level and time parameter t_u $[d^{-1}]$. When the store reaches maximum capacity S_{b2} [mm], excess flow Q_{2f} $[mm/d]$ is routed towards the fast response routing store.

$$\frac{dS_{c1}}{dt} = Q_{1f} + Q_{2f} - Q_f \tag{361}$$

$$Q_f = t_c * S_{c1} \tag{362}$$

Where S_{c1} [mm] is current storage in the fast flow routing reservoir, refilled by Q_{1f} and Q_{2f}. Routed flow Q_f depends on the mean residence time parameter t_c $[d^{-1}]$.

$$\frac{dS_{c2}}{dt} = Q_{2u} - Q_u \tag{363}$$

$$Q_u = t_c * S_{c2} \tag{364}$$

Where S_{c2} [mm] is current storage in the slow flow routing reservoir, refilled by Q_{2u}. Routed flow Q_u depends on the mean residence time parameter t_c $[d^{-1}]$. Total simulated flow Q_t $[mm/d]$:

$$Q_t = Q_f + Q_u \tag{365}$$

S2.32 MOPEX-4 (model ID: 32)

The MOPEX-4 model (fig. S33) is part of a model improvement study that investigates the relationship between dominant processes and model structures for 197 catchments in the MOPEX database (Ye et al., 2012). It has 5 stores and 10 parameters (T_{crit}, ddf, S_{b1}, t_w, I_α, I_s, S_{b2}, t_u, S_e, t_c). The original model relies on observations of Leaf Area Index and a calibrated interception fraction. Liang et al. (1994) show typical Leaf Area Index time series, and a sinusoidal function is a reasonable approximation of this. Therefore, the model is slightly modified to use a calibrated sinusoidal function, so that the data input requirements for MOPEX-4 are consistent with other models. The model aims to represent:

- Snow accumulation and melt;
- Time-varying interception;
- Saturation excess flow;
- Infiltration to deeper soil layers;
- A split between fast and slow runoff.

S2.32.1 File names

Model: m_32_mopex4_10p_5s
Parameter ranges: m_32_mopex4_10p_5s_parameter_ ranges

S2.32.2 Model equations

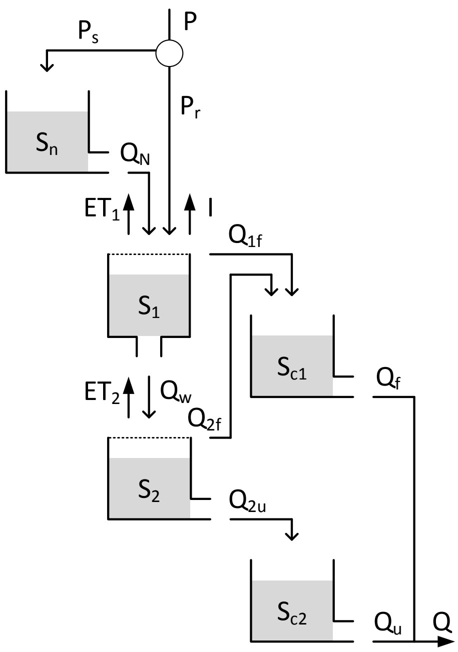

Figure S33: Structure of the MOPEX-4 model

$$\frac{dS_n}{dt} = P_s - Q_n \tag{366}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{367}$$

$$Q_n = \begin{cases} ddf * (T - T_{crit}), & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{368}$$

Where S_n [mm] is the current snow pack. Precipitation occurs as snowfall P_s $[mm/d]$ when current temperature T $[^oC]$ is below threshold T_{crit} $[^oC]$. Snowmelt Q_N $[mm/d]$ occurs when the temperature rises above the threshold temperature and relies in the degree-day factor dd $[mm/^oC/d]$.

$$\frac{dS_1}{dt} = P_r - ET_1 - I - Q_{1f} - Q_w \tag{369}$$

$$P_r = \begin{cases} P, & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{370}$$

$$ET_1 = \frac{S_1}{S_{b1}} * Ep \tag{371}$$

$$I = max\left(0, I_\alpha + (1 - I_\alpha)cos\left(2\pi\frac{t - I_s}{t_{max}}\right)\right) * P_r \tag{372}$$

$$Q_{1f} = \begin{cases} P, & \text{if } S_1 \geq S_{b1} \\ 0, & \text{otherwise} \end{cases} \tag{373}$$

$$Q_w = t_w * S_1 \tag{374}$$

Where S_1 [mm] is the current storage in soil moisture and P_r precipitation as rain $[mm/d]$. Evaporation ET_1 $[mm/d]$ depends linearly on current soil moisture, maximum soil moisture S_{b1} [mm] and potential evapotranspiration E_p [mm/d]. Interception I $[mm/d]$ depends on the mean intercepted fraction I_α [-], the maximum Leaf Area Index timing I_s [d] and the length of the seasonal cycle t_{max} [d] (usually set at 365 days). Saturation excess flow Q_{1f} $[mm/d]$ occurs when the soil moisture bucket exceeds its maximum capacity. Infiltration to deeper groundwater Q_w $[mm/d]$ depends on current soil moisture and time parameter t_w $[d^{-1}]$.

$$\frac{dS_2}{dt} = Q_w - ET_2 - Q_{2u} - Q_{2f} \tag{375}$$

$$ET_2 = \frac{S_2}{S_e} * Ep \tag{376}$$

$$Q_{2u} = t_u * S_2 \tag{377}$$

$$Q_{2f} = \begin{cases} Q_w, & \text{if } S_2 \geq S_{b2} \\ 0, & \text{otherwise} \end{cases} \tag{378}$$

Where S_2 [mm] is the current groundwater storage, refilled by infiltration from S_1. Evaporation ET_2 $[mm/d]$ depends linearly on current groundwater and root zone storage capacity S_e [mm]. Leakage to the slow runoff store Q_{2u} $[mm/d]$ depends on current groundwater level and time parameter t_u $[d^{-1}]$. When the store reaches maximum capacity S_{b2} [mm], excess flow Q_{2f} $[mm/d]$ is routed towards the fast response routing store.

$$\frac{dS_{c1}}{dt} = Q_{1f} + Q_{2f} - Q_f \tag{379}$$

$$Q_f = t_c * S_{c1} \tag{380}$$

Where S_{c1} [mm] is current storage in the fast flow routing reservoir, refilled by Q_{1f} and Q_{2f}. Routed flow Q_f depends on the mean residence time parameter t_c $[d^{-1}]$.

$$\frac{dS_{c2}}{dt} = Q_{2u} - Q_u \tag{381}$$

$$Q_u = t_c * S_{c2} \tag{382}$$

Where S_{c2} [mm] is current storage in the slow flow routing reservoir, refilled by Q_{2u}. Routed flow Q_u depends on the mean residence time parameter t_c $[d^{-1}]$. Total simulated flow Q_t $[mm/d]$:

$$Q_t = Q_f + Q_u \tag{383}$$

S2.33 SACRAMENTO model (model ID: 33)

The SACRAMENTO model (fig. S34) is part of an ongoing model development project by the National Weather Service, which started several decades ago (Burnash, 1995; National Weather Service, 2005). The documentation mentions a specific order of flux computations. For consistency with other models, here all fluxes are computed simultaneously. It has 5 stores and 13 parameters ($PCTIM$, $UZTWM$, $UZFWM$, k_{uz}, $PBASE$, $ZPERC$, $REXP$, $LZTWM$, $LZFWPM$, $LZFWSM$, $PFREE$, k_{lzp}, k_{lzs}). The model also uses several coefficients derived from the calibration parameters (Koren et al., 2000): $PBASE$ and $ZPERC$. The model aims to represent:

- Impervious and direct runoff;
- Within soil division of water storage between tension and free water;
- Surface runoff, interflow and percolation to deeper soil layers;
- Multiple baseflow processes.

S2.33.1 File names

Model: m_33_sacramento_11p_5s
Parameter ranges: m_33_sacramento_11p_5s_parameter_ ranges

S2.33.2 Model equations

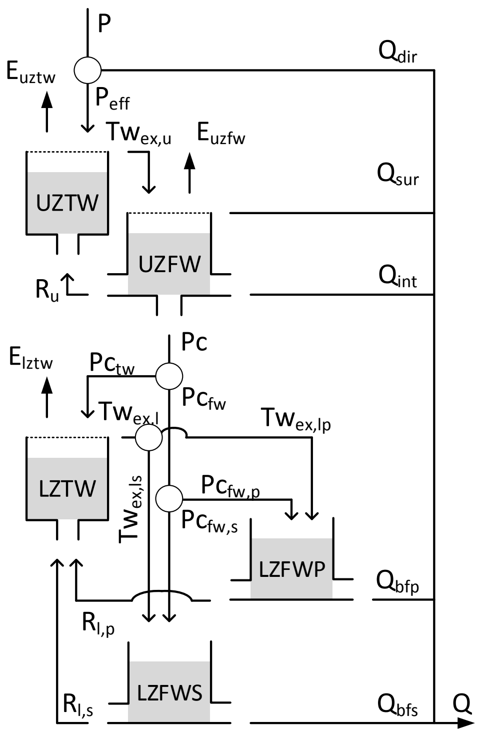

Figure S34: Structure of the SACRAMENTO model

$$\frac{dUZTW}{dt} = P_{eff} + R_u - E_{uztw} - Tw_{ex,u} \tag{384}$$

$$P_{eff} = (1 - PCTIM) * P \tag{385}$$

$$R_u = \begin{cases} \frac{UZTWM * UZFW - UZFWM * UZTW}{UZTWM + UZFWM}, & \text{if } \frac{UZTW}{UZTWM} < \frac{UZFW}{UZFWM} \\ 0, & \text{otherwise} \end{cases} \tag{386}$$

$$E_{uztw} = \frac{UZTW}{UZTWM} * E_p \tag{387}$$

$$Tw_{ex,u} = \begin{cases} P_{eff}, & \text{if } UZTW = UZTWM \\ 0, & \text{otherwise} \end{cases} \tag{388}$$

Where $UZTW$ [mm] is upper zone tension water, refilled by effective precipitation

P_{eff} $[mm/d]$ and redistribution of free water R_u $[mm/d]$, and is drained by evaporation E_{uztw} $[mm/d]$ and tension water excess $Tw_{ex,u}$ $[mm/d]$. P_{eff} is the fraction $(1-PCTIM)$ [-] of precipitation P that does not fall on impervious fraction $PCTIM$ [-]. R_u is only active when the relative deficit in tension water is greater than that in free water, and rebalances the available water in the upper zone. This uses the current storages, $UZTW$ and $UZFW$, and maximum storages, $UZTWM$ [mm] and $UZFWM$ [mm], of tension and free water stores respectively. Evaporation is determined with a linear relation between available, maximum upper zone tension storage and potential evapotranspiration E_p [mm/d]. $Tw_{ex,u}$ occurs only when the store is at maximum capacity.

$$\frac{dUZFW}{dt} = Tw_{ex,u} - E_{uzfw} - Q_{sur} - Q_{int} - Pc - R_u \tag{389}$$

$$E_{uzfw} = \begin{cases} E_p - E_{uztw}, & \text{if } UZFW > 0 \;\&\; E_p > E_{uztw} \\ 0, & \text{otherwise} \end{cases} \tag{390}$$

$$Q_{sur} = \begin{cases} Tw_{ex,u}, & \text{if } UZFW = UZFWM \\ 0, & \text{otherwise} \end{cases} \tag{391}$$

$$Q_{int} = k_{uz} * UZFW \tag{392}$$

$$Pc = Pc_{demand} * \frac{UZFW}{UZFWM} \tag{393}$$

$$Pc_{demand} = PBASE * \left(1 + ZPERC * \left(\frac{\sum LZ_{deficiency}}{\sum LZ_{capacity}}\right)^{1+REXP}\right) \tag{394}$$

$$LZ_{deficiency} = [LZTWM - LZTW] + [LZFWPM - LZWFP] + [LZFWSM - LZFWS] \tag{395}$$

$$LZ_{capacity} = LZTWM + LZFWPM + LZFWSM \tag{396}$$

Where $UZFW$ [mm] is upper zone free water, refilled by excess water $Tw_{ex,u}$ that can not be stored as tension water, and drained by evaporation E_{uzfw} $[mm/d]$, surface runoff Q_{sur} $[mm/d]$, interflow Q_{int} $[mm/d]$, and percolation to deeper groundwater Pc $[mm/d]$. Evaporative demand unmet by the upper tension water store is taken from upper free water storage at the potential rate. Q_{sur} occurs only when the store is at maximum capacity $UZFWM$ [mm]. Q_{int} uses time coefficient k_{uz} $[d^{-1}]$ to simulate interflow. Percolation Pc is calculated as a balance between the fraction water availability in upper zone free storage, and demand from the lower zone Pc_{demand}. The demand can be between a base percolation rate $PBASE$ $[mm/d]$ and an upper limit of $ZPERC$ [-] times $PBASE$. This demand is scaled by the relative size of lower zone moisture deficiencies, expressed as the ratio between total deficiency and maximum lower zone storage. $LZTWM$ [mm], $LZFWP$ [mm], $LZFWS$ [mm] are the maximum capacity of the lower zone tension store, primary free water store and supplemental free water store respectively. The lower zone percolation demand is potentially non-linear through exponent $REGX$ [-]. $PBASE$ is calculated as $k_{lzp} * LZFWPM + K_{lzs} * LZFWSM$.

$$\frac{dLZTW}{dt} = Pc_{tw} + R_{l,p} + R_{l,s} - E_{lztw} - Tw_{ex,l} \tag{397}$$

$$Pc_{tw} = (1 - PFREE) * Pc \tag{398}$$

$$R_{l,p} = \begin{cases} LZFWPM * \frac{-LZTW(LZFWPM + LZFWSM) + LZTWM(LZWFP + LZWFS)}{(LZFWPM + LZFWSM)(LZTWM + LZFWPM + LZFWSM)}, \\ \qquad \text{if } \frac{LZTW}{LZTWM} < \frac{LZFWP + LZFWS}{LZFWPM + LZFWSM} \\ 0, \quad \text{otherwise} \end{cases} \tag{399}$$

$$R_{l,s} = \begin{cases} LZFWSM * \frac{-LZTW(LZFWPM + LZFWSM) + LZTWM(LZWFP + LZWFS)}{(LZFWPM + LZFWSM)(LZTWM + LZFWPM + LZFWSM)}, \\ \qquad \text{if } \frac{LZTW}{LZTWM} < \frac{LZFWP + LZFWS}{LZFWPM + LZFWSM} \\ 0, \quad \text{otherwise} \end{cases} \tag{400}$$

$$E_{lztw} = \begin{cases} (E_p - E_{uztw} - E_{uzfw}) * \frac{LZTW}{UZTWM + LZTWM}, & \text{if } LZTW > 0 \;\&\; E_p > (E_{uztw} + E_{uzfw}) \\ 0, & \text{otherwise} \end{cases} \tag{401}$$

$$Tw_{ex,l} = \begin{cases} Pc_{tw}, & \text{if } LZTW = LZTWM \\ 0, & \text{otherwise} \end{cases} \tag{402}$$

Where $LZTW$ [mm] is lower zone tension water, refilled by percolation Pc_{tw} $[mm/d]$ and drained by evaporation E_{lztw} $[mm/d]$ and tension water excess $Tw_{ex,l}$ $[mm/d]$. Evaporative demand unmet b the upper zone can be satisfied from the lower zone tension water store, scaled by the current lower zone storage relative to total tension zone storage. Both $R_{l,p}$ and $R_{l,s}$ are only active when the relative deficit in tension water is greater than that in free water, and rebalances the available water in the lower zone. This uses the current storages, $LZTW$, $LZFWP$ and $LZFWS$, and maximum storages, $LZTWM$ [mm], $LZFWPM$ [mm] and $LZFWSM$ [mm], of the tension and free water stores respectively. Pc_{tw} is the fraction $(1 - PFREE)$ [-] of percolation Pc that does not go into free storage. $Tw_{ex,l}$ occurs only when the store is at maximum capacity $LZTWM$ [mm].

$$\frac{dLZFWP}{dt} = Pc_{fw,p} + Tw_{ex,lp} - Q_{bfp} \tag{403}$$

$$Pc_{fw,p} = \left[\frac{LZFWPM - LZFWP}{LZFWPM\left(\frac{LZFWPM-LZFWP}{LZFWPM} + \frac{LZFWSM-LZFWS}{LZFWSM}\right)} \right] * (PFREE * Pc) \tag{404}$$

$$Tw_{ex,lp} = \left[\frac{LZFWPM - LZFWP}{LZFWPM\left(\frac{LZFWPM-LZFWP}{LZFWPM} + \frac{LZFWSM-LZFWS}{LZFWSM}\right)} \right] * Tw_{ex,l} \tag{405}$$

$$Q_{bfp} = k_{lzp} * LZFWP \tag{406}$$

Where $LZFWP$ [mm] is current storage in the primary lower zone free water store, refilled by excess tension water $TW_{ex,lp}$ $[mm/d]$ and percolation $Pc_{fw,p}$ $[mm/d]$ and drained by primary baseflow Q_{bfp} $[mm/d]$. Refilling of both lower zone free water stores (primary and supplemental) is divided between the two based on their relative, scaled moisture deficiency. Percolation from the upper zone $Pc_{fw,p}$ is scaled according to the relative current moisture deficit $\frac{LZFWPM-LZFWP}{LZFWM}$ compared to the total relative deficit in the lower free water stores $\left(\frac{LZFWPM-LZFWP}{LZFWPM} + \frac{LZFWSM-LZFWS}{LZFWSM}\right)$. $Tw_{ex,lp}$ is a similarly scaled part of $Tw_{ex,l}$. Q_{bfp} uses time parameter K_{lzp} $[d^{-1}]$ to estimate primary baseflow.

$$\frac{dLZFWS}{dt} = Pc_{fw,s} + Tw_{ex,ls} - Q_{bfs} \tag{407}$$

$$Pc_{fw,s} = \left[\frac{LZFWSM - LZFWS}{LZFWSM\left(\frac{LZFWPM-LZFWP}{LZFWPM} + \frac{LZFWSM-LZFWS}{LZFWSM}\right)} \right] * (PFREE * Pc) \tag{408}$$

$$Tw_{ex,ls} = \left[\frac{LZFWSM - LZFWS}{LZFWSM\left(\frac{LZFWPM-LZFWP}{LZFWPM} + \frac{LZFWSM-LZFWS}{LZFWSM}\right)} \right] * Tw_{ex,l} \tag{409}$$

$$Q_{bfs} = k_{lzs} * LZFWS \tag{410}$$

Where $LZFWS$ [mm] is current storage in the supplemental free water lower zone store, refilled by excess tension water $TW_{ex,ls}$ $[mm/d]$ and percolation $Pc_{fw,s}$ $[mm/d]$, and drained by supplemental baseflow Q_{bfs} $[mm/d]$. $Pc_{fw,s}$ is determined based on relative deficits in the lower zone free stores, as is $Tw_{ex,ls}$. Q_{bfs} uses time parameter K_{lzs} $[d^{-1}]$ to estimate supplementary baseflow. Total simulated outflow:

$$Q_t = Q_{dir} + Q_{sur} + Q_{int} + Q_{bfp} + Q_{bfs} \tag{411}$$

S2.34 FLEX-IS (model ID: 34)

The FLEX-IS model (fig. S35) is a combination of the FLEX-B model expanded with an interception (I) routine (Fenicia et al., 2008) and a snow (S) module (Nijzink et al., 2016). It has 5 stores and 12 parameters (TT, ddf, I_{max}, UR_{max}, β, L_p, $Perc_{max}$, D, $N_{lag,f}$, $N_{lag,s}$, K_f, K_s). The model aims to represent:

- Snow accumulation and melt;
- Interception by vegetation;
- Infiltration and saturation excess flow based on a distribution of different soil depths;
- A split between fast saturation excess flow and preferential recharge to a slow store;
- Percolation from the unsaturated zone to a slow runoff store.

S2.34.1 File names

Model: m_34_flexis_12p_5s
Parameter ranges: m_34_flexis_12p_5s_parameter_ ranges

S2.34.2 Model equations

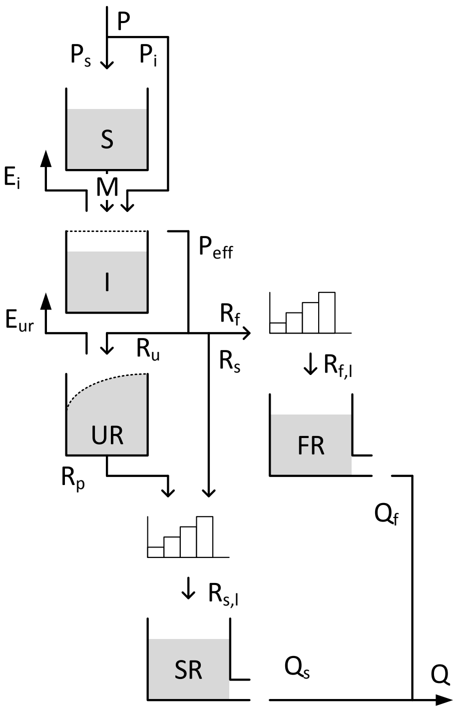

Figure S35: Structure of the FLEX-IS model

$$\frac{dS}{dt} = P_s - M \tag{412}$$

$$P_s = \begin{cases} P, & \text{if } T \leq TT \\ 0, & \text{otherwise} \end{cases} \tag{413}$$

$$M = \begin{cases} ddf * (T - TT), & \text{if } T \geq TT \\ 0, & \text{otherwise} \end{cases} \tag{414}$$

Where S [mm] is the current snow storage, P_s the precipitation that falls as snow [mm/d], M the snowmelt $[mm/d]$ based on a degree-day factor (ddf, [mm/°C/d]) and threshold temperature for snowfall and snowmelt (TT, [°C]).

$$\frac{dI}{dt} = P_I + M - E_I - P_{eff} \quad (415)$$

$$P_i = \begin{cases} P, & \text{if } T > TT \\ 0, & \text{otherwise} \end{cases} \quad (416)$$

$$E_i = \begin{cases} E_p, & \text{if } I > 0 \\ 0, & \text{otherwise} \end{cases} \quad (417)$$

$$P_{eff} = \begin{cases} P_i, & \text{if } I = I_{max} \\ 0, & \text{otherwise} \end{cases} \quad (418)$$

Where P_I $[mm/d]$ is the incoming precipitation, I is the current interception storage [mm], which is assumed to evaporate (E_i $[mm/d]$) at the potential rate E_p $[mm/d]$ when possible. When I exceeds the maximum interception storage I_{max} [mm], water is routed to the rest of the model as P_{eff} $[mm/d]$.

$$\frac{dUR}{dt} = R_u - E_{ur} - R_p \quad (419)$$

$$R_u = (1 - \Big[1 + exp\Big(\frac{-UR/UR_{max} + 1/2}{\beta}\Big)\Big]^{-1}) * P_{eff} \quad (420)$$

$$E_{ur} = E_p * min\Big(1, \frac{UR}{UR_{max}} \frac{1}{L_p}\Big) \quad (421)$$

$$R_p = Perc_{max} * \frac{-UR}{UR_{max}} \quad (422)$$

Where UR is the current storage in the unsaturated zone [mm]. R_u $[mm/d]$ is the inflow into UR based on its current storage compared to maximum storage UR_{max} [mm] and a shape distribution parameter β [-]. E_{ur} the evaporation $[mm/d]$ from UR which follows a linear relation between current and maximum storage until a threshold L_p [-] is exceeded. R_p $[mm/d]$ is the percolation from UR to the slow reservoir SR [mm], based on a maximum percolation rate $Perc_{max}$ $[mm/d]$, relative to the fraction of current storage and maximum storage.

$$R_s = (P_{eff} - R_u) * D \quad (423)$$

$$R_f = (P_{eff} - R_u) * (1 - D) \quad (424)$$

Where R_s and R_f are the flows $[mm/d]$ to the slow and fast runoff reservoir respectively, based on runoff partitioning coefficient D [-]. Both are lagged by linearly increasing triangular transformation functions with parameters $N_{lag,s}$ and $N_{lag,f}$ respectively, that give the number of time steps over which R_s and R_f need to be transformed. R_p is added to R_s before the transformation occurs.

$$\frac{dFR}{dt} = R_{f,l} - Q_f \tag{425}$$

$$Q_f = K_f * FR \tag{426}$$

Where FR is the current storage [mm] in the fast flow reservoir. Outflow Q_f $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_f $[d^{-1}]$.

$$\frac{dSR}{dt} = R_{s,l} - Q_s \tag{427}$$

$$Q_s = K_s * SR \tag{428}$$

Where SR is the current storage [mm] in the slow flow reservoir. Outflow Q_s $[mm/d]$ from the reservoir has a linear relation with storage through time scale parameter K_s $[d^{-1}]$.

$$Q = Q_f + Q_s \tag{429}$$

Where Q $[mm/d]$ is the total simulated flow as the sum of Q_s and Q_f.

S2.35 MOPEX-5 (model ID: 35)

The MOPEX-5 model (fig. S36) is part of a model improvement study that investigates the relationship between dominant processes and model structures for 197 catchments in the MOPEX database (Ye et al., 2012). It has 5 stores and 12 parameters (T_{crit}, ddf, S_{b1}, t_w, I_α, I_s, T_{min}, T_{max}, S_{b2}, t_u, S_e, t_c). The original model relies on observations of Leaf Area Index and a calibrated interception fraction. Liang et al. (1994) show typical Leaf Area Index time series, and a sinusoidal function is a reasonable approximation of this. Therefore, the model is slightly modified to use a calibrated sinusoidal function, so that the data input requirements for MOPEX-5 are consistent with other models. The model aims to represent:

- Snow accumulation and melt;
- Time-varying interception and the impact of phenology on transpiration;
- Saturation excess flow;
- Infiltration to deeper soil layers;
- A split between fast and slow runoff.

S2.35.1 File names

Model: m_35_mopex5_12p_5s
Parameter ranges: m_35_mopex5_12p_5s_parameter_ ranges

S2.35.2 Model equations

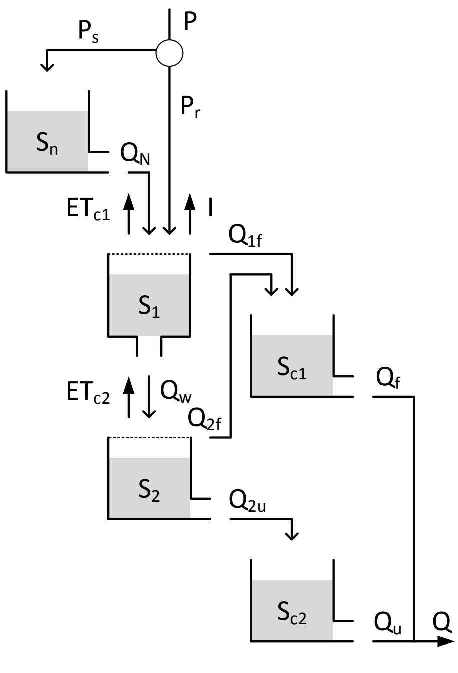

Figure S36: Structure of the MOPEX-5 model

$$\frac{dS_n}{dt} = P_s - Q_n \tag{430}$$

$$P_s = \begin{cases} P, & \text{if } T \leq T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{431}$$

$$Q_n = \begin{cases} ddf * (T - T_{crit}), & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{432}$$

Where S_n [mm] is the current snow pack. Precipitation occurs as snowfall P_s $[mm/d]$ when current temperature T $[^oC]$ is below threshold T_{crit} $[^oC]$. Snowmelt Q_N $[mm/d]$ occurs when the temperature rises above the threshold temperature and relies in the degree-day factor dd $[mm/^oC/d]$.

$$\frac{dS_1}{dt} = P_r - ET_1 - I - Q_{1f} - Q_w \tag{433}$$

$$P_r = \begin{cases} P, & \text{if } T > T_{crit} \\ 0, & \text{otherwise} \end{cases} \tag{434}$$

$$ET_{c1} = \frac{S_1}{S_{b1}} * Ep_c \tag{435}$$

$$I = max\left(0, I_\alpha + (1 - I_\alpha)sin\left(2\pi\frac{t + I_s}{365/d}\right)\right) \tag{436}$$

$$Q_{1f} = \begin{cases} P, & \text{if } S_1 \geq S_{b1} \\ 0, & \text{otherwise} \end{cases} \tag{437}$$

$$Q_w = t_w * S_1 \tag{438}$$

Where S_1 [mm] is the current storage in soil moisture and P_r precipitation as rain $[mm/d]$. Evaporation ET_1 $[mm/d]$ depends linearly on current soil moisture, maximum soil moisture S_{b1} [mm] and phenology-corrected potential evapotranspiration:

$$Ep_c = Ep * GSI \tag{439}$$

$$GSI = \begin{cases} 0, & \text{if } T < T_{min} \\ \frac{T - T_{min}}{T_{max} - T_{min}}, & \text{if } T_{min} \geq T < T_{max} \\ 1, & \text{if } T \geq T_{max} \end{cases} \tag{440}$$

Where GSI is a growing season index based on parameters T_{min} $[^oC]$ and T_{max} $[^oC]$. Interception I $[mm/d]$ depends on the mean intercepted fraction I_α [-] and the maximum Leaf Area Index timing I_s [d]. Saturation excess flow Q_{1f} $[mm/d]$ occurs when the soil moisture bucket exceeds its maximum capacity. Infiltration to deeper groundwater Q_w $[mm/d]$ depends on current soil moisture and time parameter t_w $[d^{-1}]$.

$$\frac{dS_2}{dt} = Q_w - ET_2 - Q_{2u} - Q_{2f} \tag{441}$$

$$ET_{c2} = \frac{S_2}{S_e} * Ep_c \tag{442}$$

$$Q_{2u} = t_u * S_2 \tag{443}$$

$$Q_{2f} = \begin{cases} Q_w, & \text{if } S_2 \geq S_{b2} \\ 0, & \text{otherwise} \end{cases} \tag{444}$$

Where S_2 [mm] is the current groundwater storage, refilled by infiltration from S_1. Evaporation ET_2 $[mm/d]$ depends linearly on current groundwater and root zone storage capacity S_e [mm]. Leakage to the slow runoff store Q_{2u} $[mm/d]$ depends on current groundwater level and time parameter t_u $[d^{-1}]$. When the store reaches maximum capacity S_{b2} [mm], excess flow Q_{2f} $[mm/d]$ is routed towards the fast response routing store.

$$\frac{dS_{c1}}{dt} = Q_{1f} + Q_{2f} - Q_f \tag{445}$$

$$Q_f = t_c * S_{c1} \tag{446}$$

Where S_{c1} [mm] is current storage in the fast flow routing reservoir, refilled by Q_{1f} and Q_{2f}. Routed flow Q_f depends on the mean residence time parameter t_c $[d^{-1}]$.

$$\frac{dS_{c2}}{dt} = Q_{2u} - Q_u \tag{447}$$

$$Q_u = t_c * S_{c2} \tag{448}$$

Where S_{c2} [mm] is current storage in the slow flow routing reservoir, refilled by Q_{2u}. Routed flow Q_u depends on the mean residence time parameter t_c $[d^{-1}]$. Total simulated flow Q_t $[mm/d]$:

$$Q_t = Q_f + Q_u \tag{449}$$

S2.36 MODHYDROLOG (model ID: 36)

The MODHYDROLOG model (fig. S37) is an elaborate groundwater recharge model, originally created for use in Australia (Chiew, 1990; Chiew and McMahon, 1994). It has 5 stores (I, D, SMS, GW and CH) and 15 parameters (INSC, COEFF, SQ, SMSC, SUB, CRAK, EM, DSC, ADS, MD, VCOND, DLEV, k_1, k_2 and k_3). It originally includes a routing scheme that allows linking sub-basins together, which has been removed here. The model aims to represent:

- Interception by vegetation;
- Infiltration and infiltration excess flow;
- Depression storage and delayed infiltration;
- Preferential groundwater recharge, interflow and saturation excess flow;
- Groundwater recharge resulting from filling up of soil moisture storage capacity;
- Water exchange between shallow and deep aquifers;
- Water exchange between aquifer and river channel.

S2.36.1 File names

Model: m_36_modhydrolog_15p_5s
Parameter ranges: m_36_modhydrolog_15p_5s_parameter_ ranges

S2.36.2 Model equations

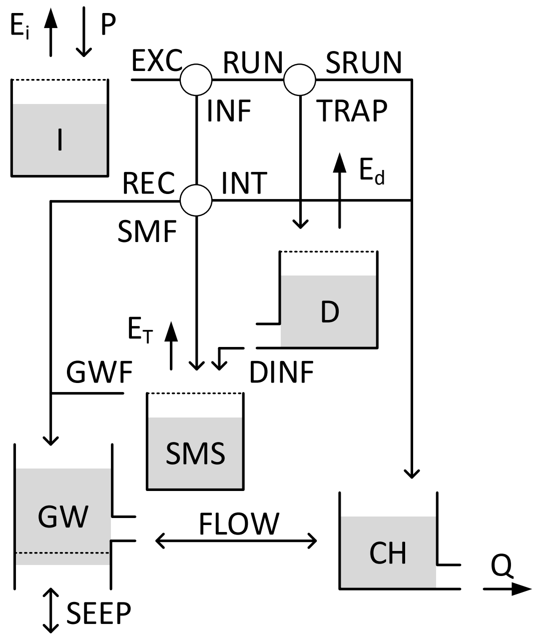

Figure S37: Structure of the MODHYDROLOG model

$$\frac{dI}{dt} = P - E_i - EXC \tag{450}$$

$$E_i = \begin{cases} E_p, & \text{if } I > 0 \\ 0, & \text{otherwise} \end{cases} \tag{451}$$

$$EXC = \begin{cases} P, & \text{if } I = INSC \\ 0, & \text{otherwise} \end{cases} \tag{452}$$

Where I [mm] is the current interception storage, P the rainfall [mm/d], E_i the evaporation from the interception store [mm/d] and EXC the excess rainfall [mm/d]). Evaporation is assumed to occur at the potential rate E_p [mm/d] when possible. When I exceeds the maximum interception capacity $INSC$, water is ruted to the rest of the model as excess precipitation EXC. The soil moisture store

SMS is instrumental in dividing runoff between infiltration and surface flow:

$$\frac{dSMS}{dt} = SMF + DINF - E_T - GWF \tag{453}$$

$$SMF = INF - INT - REC \tag{454}$$

$$INF = min\left(COEFF * exp\left(\frac{-SQ * SMS}{SMSC}\right), EXC\right) \tag{455}$$

$$INT = SUB * \frac{SMS}{SMSC} * INF \tag{456}$$

$$REC = CRAK * \frac{SMS}{SMSC} * (INF - INT) \tag{457}$$

$$E_T = min\left(EM * \frac{SMS}{SMSC}, PET\right) \tag{458}$$

$$GWF = \begin{cases} SMF, & \text{if } SMS = SMSC \\ 0, & \text{otherwise} \end{cases} \tag{459}$$

Where SMS is the current storage in the soil moisture store [mm]. SMF [mm/d] and DINF [mm/d] are the infiltration and delayed infiltration respectively. INF is total infiltration [mm/d] from excess precipitation, based on maximum infiltration loss parameter COEFF [-], the infiltration loss exponent SQ and the ratio between current soil misture storage SMS [mm] and the maximum soil moisture capacity SMSC [mm]. INT represents interflow and saturation excess flow [mm/d], using a constant of proportionality SUB [-]. REC is preferential recharge of groundwater [mm/d] based on another constant of proportionality CRAK [-]. SMF is flow into soil moisture storage [mm/d]. E_T evaporation from the soil moisture that occurs at the potential rate when possible [mm/d], based on the maximum plant-controlled rate EM [mm/d]. GWF is the flow to the groundwater store [mm/d]:

$$\frac{dD}{dt} = TRAP - E_D - DINF \tag{460}$$

$$TRAP = ADS * exp\left(-MD\frac{D}{DSC-D}\right) * RUN \tag{461}$$

$$RUN = EXC - INF \tag{462}$$

$$E_D = \begin{cases} ADS * E_p, & \text{if } D > 0 \\ 0, & \text{otherwise} \end{cases} \tag{463}$$

$$DINF = \begin{cases} ADS * RATE, & \text{if } D > 0 \\ 0, & \text{otherwise} \end{cases} \tag{464}$$

$$RATE = COEFF * exp\left(-SQ\frac{SMS}{SMSC}\right) - INF - INT - REC \tag{465}$$

Where TRAP [mm/d] is the part of overland flow captured in the depression store (equation taken from Porter and McMahon (1971)), E_D the evaporation from the depression store [mm/d], and DINF delayed infiltration to soil moisture [mm/d]. TRAP uses DSC as the maximum depression store capacity [mm], ADS as the fraction of land functioning as depression storage [-] and MD a depression storage parameter [-]. E_D relies on the potential evapotranspiration E_p. The grundwater store has no defined upper and lower boundary and instead fluctuates around a datum DLEV:

$$\frac{dGW}{dt} = REC + GWF - SEEP - FLOW \tag{466}$$

$$SEEP = VCOND * (GW - DLEV) \tag{467}$$

$$FLOW = \begin{cases} k_1 * |GW| + k_2 * (1 - exp(-k_3 * |GW|)), & \text{if } GW \geq 0 \\ -(k_1 * |GW| + k_2 * (1 - exp(-k_3 * |GW|))), & \text{if } GW < 0 \end{cases} \tag{468}$$

Where SEEP [mm/d] is the exchange with a deeper aquifer (can be negative or positive) and FLOW [mm/d] the exchange with the channel (can be negative or positive). VCOND is a leakage coefficient, DLEV a datum around which the groundwater level can fluctuate, and k_1, k_2 and k_3 are runoff coefficients. The channel store aggregates incoming fluxes and produces the total runoff Q_t [mm/d]:

$$\frac{dCH}{dt} = SRUN + INT + FLOW - Q \tag{469}$$

$$SRUN = RUN - TRAP \tag{470}$$

$$Q_t = \begin{cases} CH, & \text{if } CH > 0 \\ 0, & \text{otherwise} \end{cases} \tag{471}$$

S2.37 HBV-96 (model ID: 37)

The HBV-96 model (fig. S38) was originally developed for use in Sweden, but has been widely applied beyond its original region (Lindström et al., 1997). It can account for different land types (forest, open ground, lakes) but that distinction has been removed here. It has 5 stores and 15 parameters (TT, TTI, CFR, $CFMAX$, TTM, WHC, $CFLUX$, FC, LP, β, K_0, α, c, K_1, $MAXBAS$) parameters. The model aims to represent:

- Snow accumulation, melt and refreezing;
- Infiltration and capillary flow to, and evaporation from, soil moisture;
- A non-linear storage-runoff relationship from the upper runoff-generating zone;
- A linear storage-runoff relationship from the lower runoff-generating zone.

S2.37.1 File names

Model: m_37_hbv_15p_5s
Parameter ranges: m_37_hbv_15p_5s_parameter_ ranges

S2.37.2 Model equations

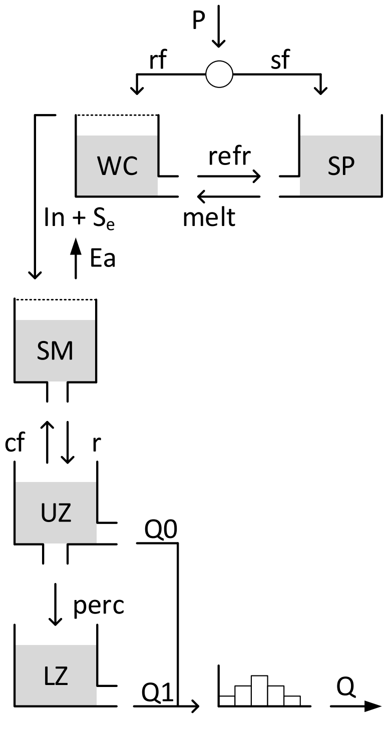

Figure S38: Structure of the HBV-96 model

$$\frac{dSP}{dt} = sf + refr - melt \tag{472}$$

$$sf = \begin{cases} P, & \text{if } T \leq TT - \frac{1}{2}TTI \\ P * \frac{TT + \frac{1}{2}TTI - T}{TTI}, & \text{otherwise} \\ 0, & \text{if } T \geq TT + \frac{1}{2}TTI \end{cases} \tag{473}$$

$$refr = \begin{cases} CFR * CFMAX * (TTM - T), & \text{if } T < TTM \\ 0, & \text{otherwise} \end{cases} \tag{474}$$

$$melt = \begin{cases} CFMAX * (T - TTM), & \text{if } T \geq TTM \\ 0, & \text{otherwise} \end{cases} \tag{475}$$

Where SP is the current snow storage [mm]. sf is precipitation that occurs as snowfall [mm/d] based on daily precipitation P [mm/d], threshold temperature for snowfall TT [°C] and the snowfall threshold interval length TTI [°C]. $refr$ [mm/d] is the refreezing of liquid snow if the current temperature T is below the melting threshold TTM [°C], using a coefficient of refreezing CFR [-] and a degree-day factor CFMAX [mm/d/°C]. *melt* represents snowmelt if the current temperature T is below the melting threshold TTM, using the degree-day factor CFMAX.

$$\frac{dWC}{dt} = rf + melt - refr - in - S_{excess} \tag{476}$$

$$rf = \begin{cases} 0, & \text{if } T \leq TT - \frac{1}{2}TTI \\ P * \frac{T-TT+\frac{1}{2}TTI}{TTI}, & \text{otherwise} \\ P, & \text{if } T \geq TT + \frac{1}{2}TTI \end{cases} \tag{477}$$

$$in = \begin{cases} rf + melt, & \text{if } WC \geq WHC * SP \\ 0, & \text{otherwise} \end{cases} \tag{478}$$

$$S_e = \begin{cases} WC - WHC * SP, & \text{if } WC \geq WHC * SP \\ 0, & \text{otherwise} \end{cases} \tag{479}$$

Where WC is the current liquid water content in the snow pack [mm], rf is the precipitation occurring as rain [mm/d] based on temperature threshold parameters TT and TTI, $refr$ is the refreezing flux, and in the infiltration to soil moisture [mm/d] that occurs when the water holding capacity of snow gets exceeded. S_{excess} [mm/d] represents excess stored water that is freed when the total possible storage of liquid water in the snow pack is reduced.

$$\frac{dSM}{dt} = (in + S_{excess}) + cf - E_a - r \tag{480}$$

$$cf = CFLUX * \left(1 - \frac{SM}{FC}\right) \tag{481}$$

$$E_a = \begin{cases} E_p, & \text{if } SM \geq LP * FC \\ E_p * \frac{SM}{LP*FC}, & \text{otherwise} \end{cases} \tag{482}$$

$$r = (in + S_{excess}) * \left(\frac{SM}{FC}\right)^{\beta} \tag{483}$$

Where SM is the current storage in soil moisture [mm], in the infiltration from the surface, cf the capillary rise [mm/d] from the unsaturated zone, E_a evaporation [mm/d] and r the flow to the upper zone [mm/d]. Capillary rise depends on the maximum rate CFLUX [mm/d], scaled by the available storage in soil moisture, expressed as the ration between current storage SM and maximum storage FC [mm]. Evaporation E_a occurs at the potential rate E_p when current soil moisture is above the wilting point LP [mm], and is scaled linearly below that. Runoff r to the upper zone has a potentially non-linear relationship with infiltration in through parameter β [-].

$$\frac{dUZ}{dt} = r - cf - Q_0 - perc \tag{484}$$
$$Q_0 = K_0 * UZ^{(1+\alpha)} \tag{485}$$
$$perc = c. \tag{486}$$

Where UZ is the current storage [mm] in the upper zone. Outflow Q_0 [mm/d] from the reservoir has a non-linear relation with storage through time scale parameter K_0 [d^{-1}] and and α [-]. Percolation *perc* [mm/d] to the lower zone is given as a constant rate c [mm/d]S.

$$\frac{dLZ}{dt} = perc - Q_1 \tag{487}$$
$$Q_1 = K_1 * LZ \tag{488}$$

Where LZ is the current storage [mm] in the lower zone. Outflow Q_1 [mm/d] from the reservoir has a linear relation with storage through time scale parameter K_1 [d^{-1}]. Total outflow is generated by summing Q_0 and Q_1 and applying a triangular transform based on lag parameter MAXBAS [d].

S2.38 Tank Model - SMA (model ID: 38)

The Tank Model (fig. S39) is originally developed for use in constantly saturated soils in Japan. This alternative Tank model - SMA (soil moisture accounting) version was developed for regions that are not continuously saturated (Sugawara, 1995). This model is identical to the original tank model, but has an increased depth in the first store to represent primary soil moisture, and adds a new store to represent secondary soil moisture. It has 5 stores and 16 parameters (sm_1, sm_2, k_1, k_2, A_0, A_1, A_2, t_1, t_2, B_0, B_1, t_3, C_0, C_1, t_4, D_1). The model aims to represent:

- Runoff on increasing time scales with depth;
- Soil moisture storage;
- capillary rise to replenish soil moisture.

S2.38.1 File names

Model: m_38_tank2_16p_5s
Parameter ranges: m_38_tank2_16p_5s_parameter_ ranges

S2.38.2 Model equations

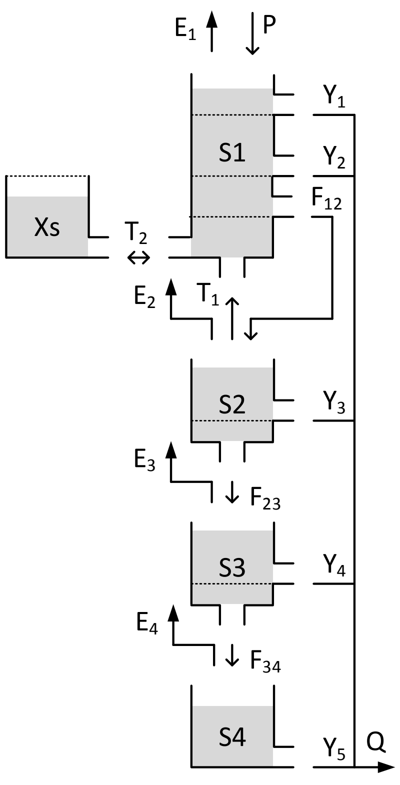

Figure S39: Structure of the Tank Model - SMA

$$\frac{dS_1}{dt} = P + T_1 - T_2 - E_1 - F_{12} - Y_2 - Y_1 \tag{489}$$

$$T_1 = k_1 \left(1 - \frac{S_1}{sm_1}\right), \text{if } S_1 < sm_1 \tag{490}$$

$$T_2 = k_2 \left(\frac{min(S_1, sm_1)}{sm_1} - \frac{X_s}{sm_2}\right) \tag{491}$$

$$E_1 = \begin{cases} Ep, & \text{if } S_1 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{492}$$

$$F_{12} = \begin{cases} A_0 * (S_1 - sm_1), & \text{if } S_1 > sm_1 \\ 0, & \text{otherwise} \end{cases} \tag{493}$$

$$Y_2 = \begin{cases} A_2 * (S_1 - t_2), & \text{if } S_1 > t_2 \\ 0, & \text{otherwise} \end{cases} \tag{494}$$

$$Y_1 = \begin{cases} A_1 * (S_1 - t_1), & \text{if } S_1 > t_1 \\ 0, & \text{otherwise} \end{cases} \tag{495}$$

Where S_1 [mm] is the current storage in the upper zone, refilled by precipitation P

$[mm/d]$ and drained by evaporation E_1 $[mm/d]$, drainage F_{12} $[mm/d]$ and surface runoff Y_1 $[mm/d]$ and Y_2 $[mm/d]$. If S_1 is below the soil moisture threshold sm_1 [mm], capillary rise T_1 $[mm/d]$ from store S_2 can occur. Capillary rise has a base rate k_1 $[mm/d]$ and decreases linearly as soil moisture S_1 nears sm_1. This store is connected to the secondary soil moisture store X_s through transfer flux T_2 $[mm/d]$. This flux can work in either direction, based on a base rate k_2 [mm/d], the current storages S_1 [mm] and X_s [mm] and the maximum soil moistures storages sm_1 [mm] and sm_2 [mm]. Evaporation E_1 occurs at the potential rate E_p $[mm/d]$ if water is available. Drainage to the intermediate layer has a linear relationship with storage through time scale parameter A_0 $[d^{-1}]$. Surface runoff Y_2 and Y_1 occur when S_1 is above thresholds t_2 [mm] and t_1 [mm] respectively. Both are linear relationships through time parameters A_2 $[d^{-1}]$ and A_1 $[d^{-1}]$ respectively.

$$\frac{dX_s}{dt} = T_2 \tag{496}$$

Where X_s [mm] is the current storage in the secondary soil moisture zone. This zone has a maximum capacity sm_2 [mm], used in the calculation of T_2. T_2 can be both positive and negative.

$$\frac{dS_2}{dt} = F_{12} - E_2 - T_1 - F_{23} - Y_3 \tag{497}$$

$$E_2 = \begin{cases} Ep, & \text{if } S_1 = 0 \ \& \ S_2 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{498}$$

$$F_{23} = B_0 * S_2 \tag{499}$$

$$Y_3 = \begin{cases} B_1 * (S_2 - t_3), & \text{if } S_2 > t_3 \\ 0, & \text{otherwise} \end{cases} \tag{500}$$

Where S_2 [mm] is the current storage in the intermediate zone, refilled by drainage F_{12} from the upper zone and drained by evaporation E_2 $[mm/d]$, drainage F_{23} $[mm/d]$ and intermediate discharge Y_3 $[mm/d]$. E_2 occurs at the potential rate E_p if water is available and the upper zone is empty. Drainage to the third layer F_{23} has a linear relationship with storage through time scale parameter B_0 $[d^{-1}]$. Intermediate runoff Y_3 occurs when S_2 is above threshold t_3 [mm] and has a linear relationship with storage through time scale parameter B_1 $[d^{-1}]$.

$$\frac{dS_3}{dt} = F_{23} - E_3 - F_{34} - Y_4 \tag{501}$$

$$E_3 = \begin{cases} Ep, & \text{if } S_1 = 0 \text{ \& } S_2 = 0 \text{ \& } S_3 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{502}$$

$$F_{34} = C_0 * S_3 \tag{503}$$

$$Y_4 = \begin{cases} C_1 * (S_3 - t_4), & \text{if } S_3 > t_4 \\ 0, & \text{otherwise} \end{cases} \tag{504}$$

Where S_3 [mm] is the current storage in the sub-base zone, refilled by drainage F_{23} from the intermediate zone and drained by evaporation E_3 $[mm/d]$, drainage F_{34} $[mm/d]$ and sub-base discharge Y_4 $[mm/d]$. E_3 occurs at the potential rate E_p if water is available and the upper zones are empty. Drainage to the fourth layer F_{34} has a linear relationship with storage through time scale parameter C_0 $[d^{-1}]$. Sub-base runoff Y_4 occurs when S_3 is above threshold t_4 [mm] and has a linear relationship with storage through time scale parameter C_1 $[d^{-1}]$.

$$\frac{dS_4}{dt} = F_{34} - E_4 - Y_5 \tag{505}$$

$$E_4 = \begin{cases} Ep, & \text{if } S_1 = 0 \text{ \& } S_2 = 0 \text{ \& } S_3 = 0 \text{ \& } S_4 > 0 \\ 0, & \text{otherwise} \end{cases} \tag{506}$$

$$Y_5 = D_1 * S_4 \tag{507}$$

Where S_4 [mm] is the current storage in the base layer, refilled by drainage F_{34} from the sub-base zone and drained by evaporation E_4 $[mm/d]$ and baseflow Y_5 $[mm/d]$. E_4 occurs at the potential rate E_p if water is available and the upper zones are empty. Baseflow Y_5 has a linear relationship with storage through time scale parameter D_1 $[d^{-1}]$. Total runoff:

$$Q_t = Y_1 + Y_2 + Y_3 + Y_4 + Y_5 \tag{508}$$

S2.39 Midlands Catchment Runoff Model (model ID: 39)

The Midlands Catchment Runoff model (fig. S40) is intended to be used in a flood-forecasting setting (Moore and Bell, 2001). To reduce the number of free parameters, the original evaporation routines and routing are somewhat simplified here. The model has 5 stores and 16 parameters (S_{max}, c_{max}, c_0, c_1, c_e, D_{surp}, k_d, γ_d, $q_{p,max}$, k_g, τ, S_{bf}, k_{cr}, γ_{cr}, k_{or} ,γ_{or}). The model aims to represent:

- Interception by vegetation;
- Direct runoff from a variable contributing area;
- A deficit-based approach to soil moisture accounting and interflow and percolation;
- Baseflow from groundwater;
- Uniform flood flood wave distribution in time;
- In-channel and out-of-channel flood routing.

S2.39.1 File names

Model: m_39_mcrm_16p_5s
Parameter ranges: m_39_mcrm_16p_5s_parameter_ ranges

S2.39.2 Model equations

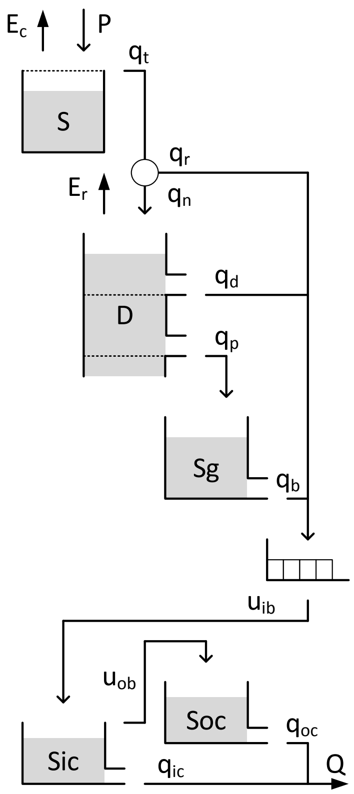

Figure S40: Structure of the MCR model

$$\frac{dS}{dt} = P - E_c - q_t \tag{509}$$

$$E_c = \begin{cases} E_p, & \text{if } S > 0 \\ 0, & \text{otherwise} \end{cases} \tag{510}$$

$$q_t = \begin{cases} P, & \text{if } S = S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{511}$$

Where S [mm] is the current interception storage, refilled by precipitation P [mm/d] and drained by evaporation E_c [mm/d] and throughfall q_t [mm/d]. E_c occurs at the potential rate whenever possible. q_t occurs only when the store is at maximum capacity S_{max} [mm].

$$\frac{dD}{dt} = q_n - E_r - q_d - q_p \tag{512}$$

$$q_n = q_t - q_r \tag{513}$$

$$q_r = min\left(c_{max}, c_0 + c_0 e^{c_1 D}\right) * q_t \tag{514}$$

$$E_r = \frac{1}{1 + e^{-c_e D}} * (E_p - E_c) \tag{515}$$

$$q_d = \begin{cases} k_d \left(D_{surp} - D\right)^{\gamma_d}, & \text{if } D > D_{surp} \\ 0, & \text{otherwise} \end{cases} \tag{516}$$

$$q_p = \begin{cases} q_{p,max}, & \text{if } D \geq D_{surp} \\ \frac{D}{D_{surp}} q_{p,max}, & \text{if } 0 < D < D_{surp} \\ 0, & \text{otherwise} \end{cases} \tag{517}$$

Where D [mm] is the current storage in soil moisture, refilled by net infiltration q_n $[mm/d]$ and drained by evaporation E_r $[mm/d]$, direct runoff q_d $[mm/d]$ and percolation q_p $[mm/d]$. Negative D-values are possible and indicate a moisture deficit. Net inflow q_n is calculated as the difference between throughfall q_t and rapid runoff q_r $[mm/d]$. q_r varies depending on the current degree of saturation in the catchment, with a maximum fraction of the catchment area contributing to rapid runoff called c_{max} [-], a minimum contributing area of c_0 [-] and an exponential increase with increasing soil moisture storage, controlled through shape parameter c_1 [-], in between. E_r fulfils any remaining evaporation demand but decreases with increasing moisture deficit (negative D values). This relation is controlled through shape parameter c_2. q_d has a non-linear relation with storage above a threshold D_{surp} [mm] through time scale parameter k_d $[d^{-1}]$ and non-linearity parameter γ_d [-]. Percolation q_p has a maximum rate of $q_{p,max}$ if D is above threshold D_{surp} and decreases linearly between $D = D_{surp}$ and $D = 0$.

$$\frac{dS_g}{dt} = q_p - q_b \tag{518}$$

$$q_b = k_g * S_g^{1.5} \tag{519}$$

Where S_g [mm] is the current groundwater storage, refilled by percolation q_p and drained by baseflow q_b $[mm/d]$. q_b uses time parameter k_b $[d^{-1}]$ and a fixed non-linearity coefficient of 1.5. Next, q_r, q_d and q_b are summed together and distributed uniformly over timespan τ [d], giving delayed flow u_{ib} $[mm/d]$.

$$\frac{dS_{ic}}{dt} = u_{ib} - u_{ob} - q_{ic} \tag{520}$$

$$u_{ob} = \begin{cases} u_{ib}, & \text{if } S_{ic} = S_{bf} \\ 0, & \text{otherwise} \end{cases} \tag{521}$$

$$q_{ic} = \begin{cases} k_{cr} * S_{ic}^{\gamma_{cr}}, & \text{if } q_{ic} < \frac{3}{4}S_{ic} \\ \frac{3}{4}S_{ic}, & \text{otherwise} \end{cases} \tag{522}$$

Where S_{ic} [mm] is the current in-channel storage, refilled by u_{ic} and drained by in-channel flow q_{ic} $[mm/d]$ and out-of-bank flow u_{ob} $[mm/d]$. u_{ob} only occurs when the store is at maximum capacity S_{bf} [mm]. q_{ic} uses time parameter k_{cr} $[d^{-1}]$ and non-linearity parameter γ_{cr} [-].

$$\frac{dS_{oc}}{dt} = u_{ob} - q_{oc} \tag{523}$$

$$q_{oc} = \begin{cases} k_{or} * S_{oc}^{\gamma_{or}}, & \text{if } q_{oc} < \frac{3}{4}S_{oc} \\ \frac{3}{4}S_{oc}, & \text{otherwise} \end{cases} \tag{524}$$

Where S_{oc} [mm] is the current out-of-channel storage, refilled by u_{ob} and drained by out-of-channel flow q_{oc} $[mm/d]$. q_{oc} uses time parameter k_{or} $[d^{-1}]$ and non-linearity parameter γ_{or} [-]. Total flow:

$$Q_t = q_{oc} + q_{ic} \tag{525}$$

S2.40 SMAR (model ID: 40)

The SMAR model (fig. S41) is the result of a series of modifications to the original 'layers-model' (O'Connell et al., 1970) and summarized by Tan and O'Connor (1996). The model uses an arbitrary number of soil moistures stores connected in series, with each store having a depth of 25mm. The number of stores is an optimization parameter. The current storage in the upper 5 stores features in various equations. For consistency within this framework, the process is reversed: the model uses a fixed number of 5 soil moisture stores, but the depth of each store is variable and given as $S_{n,max} = S_{max}/5$. It has 6 stores and 8 parameters (H, Y, S_{max}, C, G, K_G, N, K). The model aims to represent:

- Saturation excess overland flow;
- Infiltration excess overland flow;
- Gradual infiltration into soil moisture and declining evaporation potential when water is sourced from further underground;
- Groundwater flow;
- Routing of non-groundwater flow.

S2.40.1 File names

Model: m_40_smar_8p_6s
Parameter ranges: m_40_smar_8p_6s_parameter_ ranges

S2.40.2 Model equations

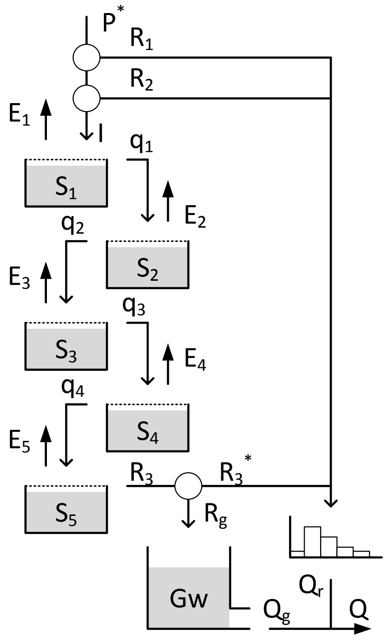

Figure S41: Structure of the SMAR model

$$\frac{dS_1}{dt} = I - E_1 - q_1 \tag{526}$$

$$I = \begin{cases} Y, & \text{if } P^* - R_1 \geq Y \\ P^* - R_1, & \text{otherwise} \end{cases} \tag{527}$$

$$P^* = \begin{cases} P - E_p, & \text{if } P > E_p \\ 0, & \text{otherwise} \end{cases} \tag{528}$$

$$R_1 = P^* * H * \frac{\sum S_n}{S_{max}} \tag{529}$$

$$R_2 = (P^* - R_1) - I \tag{530}$$

$$E_1 = C^{(1-1)} * Ep^* \tag{531}$$

$$E_p^* = \begin{cases} E_p - P, & \text{if } E_p > P \\ 0, & \text{otherwise} \end{cases} \tag{532}$$

$$q_1 = \begin{cases} P^* - R_1 - R_2, & \text{if } S_1 \geq \frac{S_{max}}{5} \\ 0, & \text{otherwise} \end{cases} \tag{533}$$

Where S_1 [mm] is the current storage in the upper soil layer, I $[mm/d]$ infiltration into the soil, P^* the effective precipitation $[mm/d]$, R_1 $[mm/d]$ is direct runoff, R_2 $[mm/d]$ is infiltration excess runoff, E_1 $[mm/d]$ evaporation and q_1 $[mm/d]$ flow towards deeper soil layers. I uses a constant infiltration rate Y $[mm/d]$. Direct runoff R_1 relies on distribution parameter H [-] and is scaled by the current soil moisture storage in all layers compared to the maximum soil moisture storage S_{max} [mm] of all layers. Evaporation from this soil layer occurs at the effect potential rate E_p^*. Runoff to deeper layers q_1 only occurs when the current storage exceeds the store's maximum capacity.

$$S_2 = q_1 - E_2 - q_2 \tag{534}$$

$$E_2 = \begin{cases} C^{(2-1)} * Ep, & \text{if } S_1 = 0 \\ 0, & \text{otherwise} \end{cases} \tag{535}$$

$$q_2 = \begin{cases} q1, & \text{if } S_2 \geq \frac{S_{max}}{5} \\ 0, & \text{otherwise} \end{cases} \tag{536}$$

Where S_2 [mm] is the current storage in the second soil layer, E_2 $[mm/d]$ the evaporation scaled by parameter C [-], and q_2 $[mm/d]$ overflow into the next layer. Evaporation is assumed to occur only when the storage in the upper layers has been exhausted.

$$S_3 = q_2 - E_3 - q_3 \tag{537}$$

$$E_3 = \begin{cases} C^{(3-1)} * Ep, & \text{if } S_2 = 0 \\ 0, & \text{otherwise} \end{cases} \tag{538}$$

$$q_3 = \begin{cases} q2, & \text{if } S_3 \geq \frac{S_{max}}{5} \\ 0, & \text{otherwise} \end{cases} \tag{539}$$

Where S_3 [mm] is the current storage in the second soil layer, E_3 $[mm/d]$ the evaporation scaled by parameter C^2 [-], and q_3 $[mm/d]$ overflow into the next layer. Evaporation is assumed to occur only when the storage in the upper layers has been exhausted.

$$S_4 = q_3 - E_4 - q_4 \tag{540}$$
$$E_4 = \begin{cases} C^{(4-1)} * Ep, & \text{if } S_3 = 0 \\ 0, & \text{otherwise} \end{cases} \tag{541}$$
$$q_4 = \begin{cases} q3, & \text{if } S_4 \geq \frac{S_{max}}{5} \\ 0, & \text{otherwise} \end{cases} \tag{542}$$

Where S_4 [mm] is the current storage in the second soil layer, E_4 $[mm/d]$ the evaporation scaled by parameter C^3 [-], and q_4 $[mm/d]$ overflow into the next layer. Evaporation is assumed to occur only when the storage in the upper layers has been exhausted.

$$S_5 = q_4 - E_5 - R_3 \tag{543}$$
$$E_5 = \begin{cases} C^{(5-1)} * Ep, & \text{if } S_4 = 0 \\ 0, & \text{otherwise} \end{cases} \tag{544}$$
$$R_3 = \begin{cases} q4, & \text{if } S_5 \geq \frac{S_{max}}{5} \\ 0, & \text{otherwise} \end{cases} \tag{545}$$

Where S_5 [mm] is the current storage in the second soil layer, E_5 $[mm/d]$ the evaporation scaled by parameter C^4 [-], and R_3 $[mm/d]$ overflow towards groundwater. Evaporation is assumed to occur only when the storage in the upper layers has been exhausted.

$$\frac{dG_w}{dt} = R_g - Q_g \tag{546}$$
$$R_g = G * R_3 \tag{547}$$
$$Q_g = K_G * G_w \tag{548}$$

Where G_w [mm] is the current groundwater storage, refilled by fraction G [-] of R_3 $[mm/d]$ and drained as a linear reservoir with time parameter K_G $[d^{-1}]$. This groundwater flow Q_g $[mm/d]$ contributes directly to simulated streamflow Q. The fraction $R_3^* = (1 - G) * R_3$ that does not reach the groundwater reservoir is combined with R_1 and R_2 and routed with a gamma function with parameters N and K. The routing function approximates a Nash-cascade consisting of N reservoirs with storage coefficient K:

$$h = \frac{1}{K\Gamma(N)} \left(\frac{t}{K}\right)^{N-1} e^{-t/K} \tag{549}$$

Integration over the time step length d provides the fraction of flow routed per time step Q_r $[mm/d]$. Total flow:

$$Q_t = Q_r + Q_g \tag{550}$$

S2.41 NAM model (model ID: 41)

The NAM model (fig. S42) is originally developed for use in Denmark (Nielsen and Hansen, 1973). Here a small modification is made by replacing runoff routing equations of the form $\frac{1}{k}e^{-t/k}$ with the linear reservoirs these equations represent. The model has 6 stores and 10 parameters (C_s, C_{if}, L^*, C_{L1}, U^*, C_{of}, C_{L2}, K_0, K_1, K_b). The model aims to represent:

- Snow accumulation and melt;
- Interflow when total soil moisture exceeds a threshold;
- Separation of saturation excess flow into overland flow and infiltration;
- Baseflow from groundwater.;

S2.41.1 File names

Model:	m_41_nam_10p_6s
Parameter ranges:	m_41_nam_10p_6s_parameter_ ranges

S2.41.2 Model equations

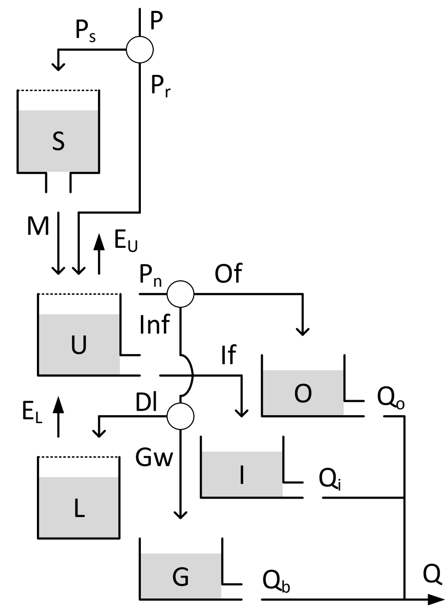

Figure S42: Structure of the NAM model

$$\frac{dS}{dt} = P_s - M \tag{551}$$

$$P_s = \begin{cases} P, & \text{if } T \leq 0 \\ 0, & \text{otherwise} \end{cases} \tag{552}$$

$$M = \begin{cases} C_s * T, & \text{if } T > 0 \\ 0, & \text{otherwise} \end{cases} \tag{553}$$

Where S is the current snow storage [mm], P_s $[mm/d]$ the precipitation that falls as snow and M the snowmelt $[mm/d]$ based on a degree-day factor (c_s, [mm/°C/d]). The freezing point of 0^o [C] is used as a threshold for snowfall and melt.

$$\frac{dU}{dt} = P_r + M - E_U - If - P_n \tag{554}$$

$$P_r = \begin{cases} P, & \text{if } T > 0 \\ 0, & \text{otherwise} \end{cases} \tag{555}$$

$$E_U = \begin{cases} E_p, & \text{if } U > 0 \\ 0, & \text{otherwise} \end{cases} \tag{556}$$

$$If = \begin{cases} C_{if} * \frac{L/L^* - C_{L1}}{1 - C_{L1}} U, & \text{if } L/L^* > C_{L1} \\ 0, & \text{otherwise} \end{cases} \tag{557}$$

$$P_n = \begin{cases} (P_r + M), & \text{if } U = U^* \\ 0, & \text{otherwise} \end{cases} \tag{558}$$

Where U [mm] is the current storage in the upper zone, refilled by precipitation as rain P_r $[mm/d]$ and snowmelt M, and drained by evaporation E_U $[mm/d]$, interflow If $[mm/d]$ and net precipitation P_n $[mm/d]$. P_r occurs only when the current temperature exceeds the threshold of 0^oC. E_U occurs at the potential rate E_p whenever possible. If occurs only if the fractional storage in the lower zone L/L^* (L is current lower zone storage, L^* is lower zone maximum storage) exceeds a threshold C_{L1} [-]. If is further scaled current deficit in the lower zone and a second scaling coefficient C_{if} [-]. P_n occurs only when the upper zone exceeds its maximum storage capacity U^* [mm].

$$\frac{dL}{dt} = Dl - E_t \tag{559}$$

$$Dl = (P_n - Of)\left(1 - \frac{L}{L^*}\right) \tag{560}$$

$$Of = \begin{cases} C_{of} * \frac{L/L^* - C_{L2}}{1 - C_{L2}} * P_n, & \text{if } L/L^* > C_{L2} \\ 0, & \text{otherwise} \end{cases} \tag{561}$$

$$E_t = \begin{cases} \frac{L}{L^*} E_p, & \text{if } U = 0 \\ 0, & \text{otherwise} \end{cases} \tag{562}$$

Where L [mm] is the current storage in the lower zone, refilled by a fraction of infiltration Dl $[mm/d]$ and drained by evaporation E_t $[mm/d]$. Dl is calculated as a fraction of infiltration $P_n - Of$, dependent on the current deficit in the lower zone. Note that with the current formulation Dl might be larger than the lower zone deficit $L^* - L$ and a constraint of the form $Dl \leq L^* - L$ is needed. Overland flow Of $[mm/d]$ is a fraction of P_n determined using the relative storage in the lower zone L/L^* and two coefficients C_{of} [-] and C_{L2} [-]. E_t occurs only when the upper zone is empty, and at a reduced rate that uses the relative storage in the lower zone.

$$\frac{dO}{dt} = Of - Q_o \tag{563}$$

$$Q_o = K_0 * O \tag{564}$$

Where O [mm] is the current storage in the overland flow routing store. Q_o is the routed overland flow, using time coefficient K_0 $[d^{-1}]$.

$$\frac{dI}{dt} = If - Q_i \tag{565}$$

$$Q_i = K_1 * I \tag{566}$$

Where I [mm] is the current storage in the interflow routing store. Q_i is the routed interflow, using time coefficient K_1 $[d^{-1}]$.

$$\frac{dG}{dt} = Gw - Q_b \tag{567}$$

$$Gw = (P_n - Of)\left(\frac{L}{L^*}\right) \tag{568}$$

$$Q_b = K_b * O \tag{569}$$

Where G [mm] is the current storage in the overland flow routing store, refilled by groundwater flow Gw $[mm/d]$. Q_b is the routed baseflow, using time coefficient K_b $[d^{-1}]$. Total flow:

$$Q = Q_o + Q_i + Q_b \tag{570}$$

S2.42 HYCYMODEL (model ID: 42)

The HYCYMODEL (fig. S43) is originally developed for use in heavily forested catchments in Japan (Fukushima, 1988). The original model specifies evaporation from the S_b store as $E_T = e_p(i) * Q_b/Q_{bc}$, if $S_u < 0$ & $S_b < S_{bc}$, with $Q_{bc} = f(S_{bc})$. However, no further details are given and S_{bc} is not listed as a parameter. We assume that S_{bc} [mm] is a threshold parameter and that evaporation potential declines linearly to zero when the store drops under this threshold. The model has 6 stores and 12 parameters (C, $I_{1,max}$, α, $I_{2,max}$, k_{in}, D_{50}, D_{16}, S_{bc}, k_b, p_b, k_h, k_c). The model aims to represent:

- Split between channel and ground precipitation;
- Interception by canopy and stems/trunks;
- Overland flow from a variable contributing area;
- Non-linear channel flow, hillslope flow and baseflow;
- Channel evaporation.

S2.42.1 File names

Model: m_42_hycymodel_12p_6s
Parameter ranges: m_42_hycymodel_12p_6s_parameter_ ranges

S2.42.2 Model equations

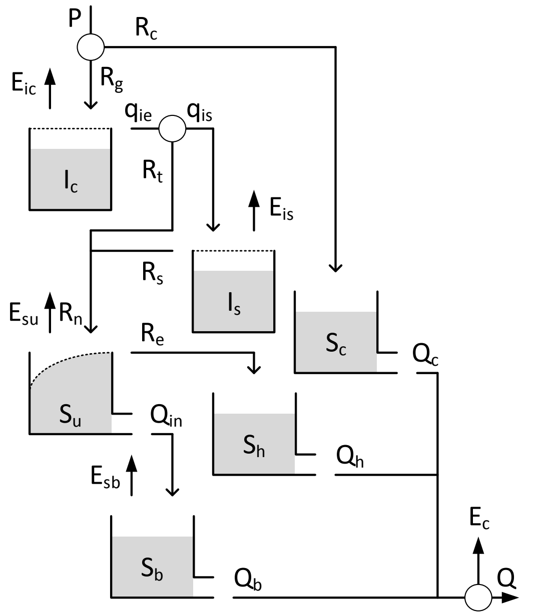

Figure S43: Structure of the HYCYMODEL

$$\frac{dI_c}{dt} = R_g - E_{ic} - q_{ie} \tag{571}$$

$$R_g = (1 - C)P \tag{572}$$

$$E_{ic} = \begin{cases} (1 - C) * E_p, & \text{if } I_c > 0 \\ 0, & \text{otherwise} \end{cases} \tag{573}$$

$$q_{ie} = \begin{cases} R_g, & \text{if } I_c = I_{1,max} \\ 0, & \text{otherwise} \end{cases} \tag{574}$$

Where I_c [mm] is the current canopy storage, refilled by rainfall on ground R_g $[mm/d]$ and drained by evaporation E_{ic} $[mm/d]$ and canopy interception excess Q_{ie} $[mm/d]$. R_g is the fraction (1-C) [mm] of rainfall P $[mm/d]$ that falls on ground (and not in the channel). This fraction appears several times in the model to scale evaporation values according to surface area.

E_{ic} occurs at the potential rate E_p $[mm/d]$ when possible. q_{ie} only occurs when the canopy store is at maximum capacity $I_{1,max}$ [mm].

$$\frac{dI_s}{dt} = q_{is} - E_{is} - R_s \tag{575}$$

$$q_{is} = \alpha * q_{ie} \tag{576}$$

$$E_{is} = \begin{cases} (1-C) * E_p, & \text{if } I_s > 0 \\ 0, & \text{otherwise} \end{cases} \tag{577}$$

$$R_s = \begin{cases} q_{is}, & \text{if } I_s = I_{2,max} \\ 0, & \text{otherwise} \end{cases} \tag{578}$$

Where I_s [mm] is the current stem and trunk storage, refilled by a fraction of canopy excess q_{is} $[mm/d]$ and drained by evaporation E_{is} $[mm/d]$ and stem flow R_s $[mm/d]$. q_{is} is the fraction α [-] of canopy excess q_{ie}. The remainder $(1-\alpha)$ is throughfall R_t $[mm/d]$. E_{is} occurs at the potential rate E_p when possible. R_s occurs only when the store is at maximum capacity $I_{2,max}$.

$$\frac{dS_u}{dt} = R_n - Re - E_{su} - Q_{in} \tag{579}$$

$$R_n = R_t + R_s \tag{580}$$

$$Re = m * R_n \tag{581}$$

$$m = \int_{-\inf}^{\xi} \frac{1}{\sqrt{2\pi}} exp\left(-\frac{\xi^2}{2}\right) d\xi \tag{582}$$

$$\xi = \frac{log\,(S_u/D_{50})}{log\,(D_{50}/D_{16})} \tag{583}$$

$$E_{su} = \begin{cases} (1-C) * E_p, & \text{if } E_{us} > 0 \\ 0, & \text{otherwise} \end{cases} \tag{584}$$

$$Q_{in} = k_{in} * S_u \tag{585}$$

Where S_u [mm] is the current storage in the upper zone, refilled by net precipitation R_n $[mm/d]$ and drained by effective rainfall R_e $[mm/d]$, evaporation E_{su} $[mm/d]$ and infiltration Q_{in} $[mm/d]$. R_n is the sum of throughfall R_t and stem flow R_s. R_e is a fraction m [-] of R_e, determined from a variable contributing area concept. m is calculated is an integral from a regular normal distribution, scaled by the current storage S_u compared to two parameters D_{50} [mm] and D_{16} [mm]. These parameters represent the effective soil depths at which respectively 50% and 16% of the catchment area contribute to R_e. E_{su} occurs at the potential rate E_p when possible. Q_{in} has a linear relation with storage through time parameter k_{in} $[d^{-1}]$.

$$\frac{dS_b}{dt} = Q_{in} - E_{sb} - Q_b \tag{586}$$

$$E_{sb} = \begin{cases} (1-C) * E_p, & \text{if } S_u = 0 \text{ \& } S_b \geq S_{bc} \\ (1-C) * E_p \frac{S_b}{S_{bc}}, & \text{otherwise} \end{cases} \tag{587}$$

$$Q_b = k_b * S_b^{p_b} \tag{588}$$

Where S_b [mm] is the current storage in the lower zone, refilled by infiltration Q_{in} and drained by evaporation E_{sb} $[mm/d]$ and baseflow Q_b $[mm/d]$. E_{sb} occurs at the potential rate when the store is above a threshold S_{bc} [mm], and declines linearly below that. Q_b has a potentially non-linear relation with storage through time parameter k_b $[d^{-1}]$ and scale parameter p_b [-].

$$\frac{dS_h}{dt} = R_e - Q_h \tag{589}$$

$$Q_h = k_h * S_h^{p_h} \tag{590}$$

Where S_h [mm] is the current storage in the hillslope routing store, refilled by effective rainfall R_e and drained by hillslope runoff Q_h. Q_h has a potentially non-linear relation with storage through time parameter k_h $[d^{-1}]$ and scale parameter p_h [-]. p_h is a fixed parameter in the original model with value 5/3.

$$\frac{dS_c}{dt} = R_c - Q_c \tag{591}$$

$$Q_c = k_c * S_c^{p_c} \tag{592}$$

Where S_c [mm] is the current storage in the channel routing store, refilled by rainfall on the channel R_c and drained by channel runoff Q_c. Q_c has a potentially non-linear relation with storage through time parameter k_c $[d^{-1}]$ and scale parameter p_c [-]. p_c is a fixed parameter in the original model with value 5/3.

$$Q_t = Q_c + Q_h + Q_b - E_c \tag{593}$$

$$E_c = C * E_p \tag{594}$$

Where Q_t $[mm/d]$ is the total flow as sum of the three individual flow fluxes minus channel evaporation E_c $[mm/d]$.

S2.43 GSM-SOCONT model (model ID: 43)

The Glacier and SnowMelt - SOil CONTribution model (GSM-SOCONT) model (fig. S44) is a model developed for alpine, partly glaciated catchments (Schaefli et al., 2005). For consistency with other models in this framework, several simplifications are used. The model does not use different elevation bands nor DEM data to estimate certain parameters, and does not calculate an annual glacier mass balance. The model has 6 stores and 12 parameters (f_{ice}, T_0, a_{snow}, T_m, k_s, a_{ice}, k_i, A, x, y, k_{sl}, β). The model aims to represent:

- Separate treatment of glacier and non-glacier catchment area;
- Snow accumulation and melt;
- Glacier melt;
- Soil moisture accounting in the non-glacier catchment area.

S2.43.1 File names

Model: m_43_gsmsocont_12p_6s
Parameter ranges: m_43_gsmsocont_12p_6s_parameter_ ranges

S2.43.2 Model equations

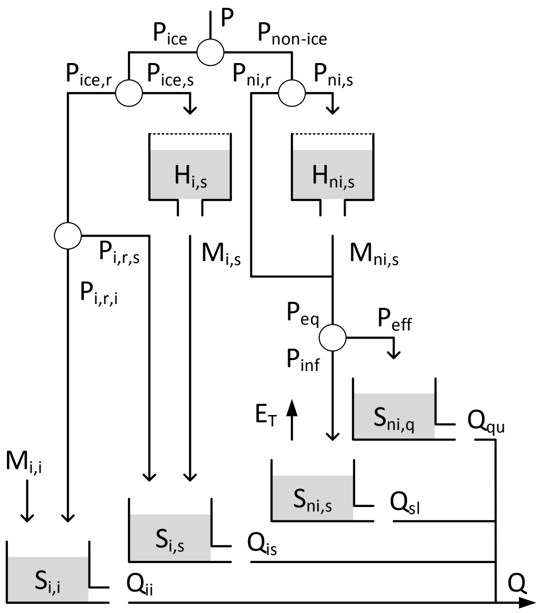

Figure S44: Structure of the GSM-SOCONT model

$$\frac{dH_{i,s}}{dt} = P_{ice,s} - M_{i,s} \tag{595}$$

$$P_{ice,s} = \begin{cases} P_{ice}, & \text{if } T \leq T_0 \\ 0, & \text{otherwise} \end{cases} \tag{596}$$

$$P_{ice} = f_{ice} * P \tag{597}$$

$$M_{i,s} = \begin{cases} a_{snow}(T - T_m), & \text{if } T > T_m \\ 0, & \text{otherwise} \end{cases} \tag{598}$$

Where $H_{i,s}$ [mm] is the current storage in the snow pack, refilled by precipitation-as-snow $P_{ice,s}$ $[mm/d]$ and depleted by melt $M_{i,s}$ $[mm/d]$. $P_{ice,s}$ occurs only when the temperature T $[^oC]$ is below a threshold temperature for snowfall T_0 $[^oC]$. P_{ice} is the fraction f_{ice} [-] of precipitation P $[mm/d]$ that falls on the ice-covered part of the catchment. $M_{i,s}$ uses a degree-day-factor a_{snow}

$[mm/^oC/d]$ to estimate snow melt if temperature is above a threshold for snow melt T_s $[^oC]$.

$$\frac{dS_{i,s}}{dt} = M_{i,s} + P_{i,r,s}, - Q_{is} \tag{599}$$

$$P_{i,r,s} = P_{ice,r}, \quad \text{if } H_{i,s} > 0 \tag{600}$$

$$P_{ice,r} = \begin{cases} P_{ice}, & \text{if } T > T_0 \\ 0, & \text{otherwise} \end{cases} \tag{601}$$

$$Q_{is} = k_s * S_{i,s} \tag{602}$$

Where $S_{i,s}$ [mm] is the current storage in the snow-water routing reservoir, refilled by snow melt $M_{i,s}$ $[mm/d]$ and rain-on-snow $P_{ice,s}$ $[mm/d]$, and drained by runoff Q_{is}. $P_{i,r,s}$ occurs only if the current snow pack storage is above zero. $P_{ice,r}$ is precipitation-as-rain that occurs only if the temperature is above a snowfall threshold T_0. Q_{is} has a linear relation with storage through time parameter k_s $[d^{-1}]$.

$$\frac{dS_{i,i}}{dt} = M_{i,i} + P_{i,r,i}, - Q_{ii} \tag{603}$$

$$P_{i,r,i} = P_{ice,r}, \quad \text{if } H_{i,s} = 0 \tag{604}$$

$$M_{i,s} = \begin{cases} a_{ice}(T - T_m), & \text{if } T > T_m \text{ \& } H_{i,s} = 0 \\ 0, & \text{otherwise} \end{cases} \tag{605}$$

$$Q_{ii} = k_i * S_{i,i} \tag{606}$$

Where $S_{i,i}$ [mm] is the current storage in the ice-water routing reservoir, refilled by glacier melt $M_{i,i}$ $[mm/d]$ and rain-on-ice $P_{ice,i}$ $[mm/d]$, and drained by runoff Q_{ii} $[mm/d]$. Both $M_{i,i}$ and $P_{ice,i}$ are assumed to only occur once the snow pack $H_{i,s}$ is depleted. $M_{i,i}$ uses a degree-day-factor a_{ice} $[mm/^oC/d]$ to estimate glacier melt. Ice storage in the glacier is assumed to be infinite. $P_{ice,r,i}$ is equal to $P_{ice,r}$ if $H_{i,s} = 0$. Q_{ii} has a linear relation with storage through time parameter k_i $[d^{-1}]$.

$$\frac{dH_{ni,s}}{dt} = P_{ni,s} - M_{ni,s} \tag{607}$$

$$P_{ni,s} = \begin{cases} P_{non-ice}, & \text{if } T \leq T_0 \\ 0, & \text{otherwise} \end{cases} \tag{608}$$

$$P_{non-ice} = (1 - f_{ice}) * P \tag{609}$$

$$M_{ni,s} = \begin{cases} a_{snow}(T - T_m), & \text{if } T > T_m \\ 0, & \text{otherwise} \end{cases} \tag{610}$$

Where $H_{ni,s}$ [mm] is the current snow pack storage on the non-ice covered fraction $1 - f_{ice}$ [-] of the catchment, which increases through snowfall $P_{ni,s}$ $[mm/d]$ and decreases through snow melt $M_{ni,s}$ $[mm/d]$. Both fluxes are calculated in the same manner as those on the ice-covered part of the catchment (fluxes $P_{ice,s}$ and $M_{ice,s}$).

$$\frac{dS_{ni,s}}{dt} = P_{inf} - E_T - Q_{sl} \tag{611}$$

$$P_{inf} = P_{eq} - P_{eff} \tag{612}$$

$$P_{eff} = P_{eq}\left(\frac{S_{ni,s}}{A}\right)^y \tag{613}$$

$$P_{eq} = M_{ni,s} + P_{ni,r} \tag{614}$$

$$E_T = E_p\left(\frac{S_{ni,s}}{A}\right)^x \tag{615}$$

$$Q_{sl} = k_{sl}S_{ni,s} \tag{616}$$

Where $S_{ni,s}$ [mm] is the current storage in soil moisture, refilled by infiltrated precipitation P_{inf} $[mm/d]$ and drained by evapotranspiration E_T $[mm/d]$ and slow flow Q_{sl} $[mm/d]$. P_{inf} depends on the effective precipitation P_{eff}. P_{eq} is the total of snow melt $M_{ni,s}$ and precipitation-as-rain $P_{ni,r}$ $[mm/d]$. $P_{ni,r}$ is calculated in the same manner as $P_{i,r}$ (equation 7). E_T is a fraction potential evapotranspiration E_p $[mm/d]$, calculated using A and non-linearity parameter y [-]. Q_{sl} has a linear relation with storage through time parameter k_{sl} $[d^{-1}]$.

$$\frac{dS_{ni,q}}{dt} = P_{eff} - Q_{qu} \tag{617}$$

$$Q_{qu} = \beta S_{ni,q}^{5/3} \tag{618}$$

Where $S_{ni,q}$ [mm] is the current storage in the direct runoff reservoir, refilled by effective precipitation P_{eff} $[mm/d]$ and by quick flow Q_{qu} $[mm/d]$. Q_{sl} has a non-linear relation with storage through time parameter β $[mm^{4/3}/d]$ and the factor 5/3. Total flow:

$$Q = Q_{qu} + Q_{sl} + Q_{is} + Q_{ii} \tag{619}$$

S2.44 ECHO model (model ID: 44)

The ECHO model (fig. S45) is a single element from the Spatially Explicit Hydrologic Response (SEHR-ECHO) model (Schaefli et al., 2014). Because the model is used as a lumped model here, the "SEHR" prefix was dropped intentionally. For consistency with other models, soil moisture storage S is given here in absolute terms [mm], rather than fractional terms that are used in the original reference. Rain- and snowfall equations are taken from Schaefli et al. (2005). The model has 6 stores and 16 parameters (ρ, T_s, T_m, a_s, a_f, G_{max}, θ, ϕ, S_{max}, sw, sm, K_{sat}, c, L_{max}, k_f, k_s). The model aims to represent:

- Interception by vegetation;
- Snowfall, snowmelt, ground-heat flux and storage and refreezing of liquid snow;
- Infiltration, infiltration excess and saturation excess;
- Fast and slow runoff.

S2.44.1 File names

Model: m_44_echo_16p_6s
Parameter ranges: m_44_echo_16p_6s_parameter_ ranges

S2.44.2 Model equations

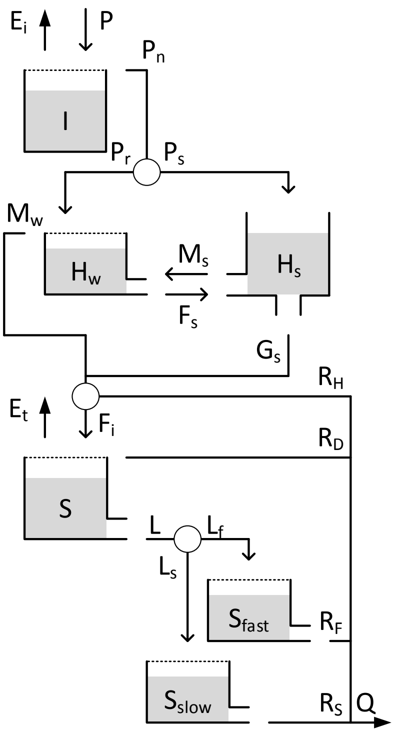

Figure S45: Structure of the ECHO model

$$\frac{dI}{dt} = P - E_i - P_n \tag{620}$$

$$E_i = \begin{cases} E_p, & \text{if } I > 0 \\ 0, & \text{otherwise} \end{cases} \tag{621}$$

$$P_n = \begin{cases} P, & \text{if } I = \rho \\ 0, & \text{otherwise} \end{cases} \tag{622}$$

Where I [mm] is the current interception storage, refilled by precipitation P [mm/d] and drained by evaporation E_i [mm/d] and net precipitation P_n [mm/d]. E_i occurs at the potential rate E_p [mm/d] when possible. P_n only occurs when the store is at maximum capacity ρ [mm].

$$\frac{dH_s}{dt} = P_s + F_s - M_s - G_s \tag{623}$$

$$P_s = \begin{cases} P_n, & \text{if } T \leq T_s \\ 0, & \text{otherwise} \end{cases} \tag{624}$$

$$M_s = \begin{cases} a_s(T - T_m), & \text{if } T > T_m, H_s > 0 \\ 0, & \text{otherwise} \end{cases} \tag{625}$$

$$F_s = \begin{cases} a_f a_s(T_m - T), & \text{if } T < T_m, H_w > 0 \\ 0, & \text{otherwise} \end{cases} \tag{626}$$

$$G_s = \begin{cases} G_{max}, & \text{if } H_s > 0 \\ 0, & \text{otherwise} \end{cases} \tag{627}$$

Where H_s [mm] is the current storage in the snow pack, refilled by precipitation-as-snow P_s $[mm/d]$ and refreezing of melted snow F_s $[mm/d]$, and drained by snowmelt M_s $[mm/d]$ and the ground-heat flux G_s $[mm/d]$. P_s is calculated as all effective rainfall after interception, provided the temperature is below a threshold T_s $[^oC]$. M_s uses a degree-day factor a_s $[mm/^oC/d]$ and threshold temperature for snowmelt T_m $[^oC]$. F_s occurs if the current temperature is below T_m and the degree-day rate reduced by factor a_f [-]. G_s occurs at a constant rate G_{max} $[mm/d]$.

$$\frac{dH_w}{dt} = P_r + M_s - F_s - M_w \tag{628}$$

$$P_r = \begin{cases} P_n, & \text{if } T > T_s \\ 0, & \text{otherwise} \end{cases} \tag{629}$$

$$M_w = \begin{cases} P_r + M_s, & \text{if } H_w = \theta * H_s \\ 0, & \text{otherwise} \end{cases} \tag{630}$$

Where H_w [mm] is the current storage of liquid water in the snow pack, refilled by precipitation-as-rain P_r $[mm/d]$ and snowmelt M_s $[mm/d]$, and drained by refreezing F_s $[mm/d]$ and outflow of melt water M_w $[mm/d]$. P_r is calculated as all effective rainfall after interception, provided the temperature is above a threshold T_s $[^oC]$. M_w occurs only if the store is at maximum capacity, which is a fraction θ [-] of the current snow pack height H_s [mm].

$$\frac{dS}{dt} = F_i - R_D - E_t - L \tag{631}$$
$$F_i = P_{eq} - R_H \tag{632}$$
$$P_{eq} = M_w + G_s \tag{633}$$
$$R_H = \begin{cases} max(P_{eq} - \phi, 0), & \text{if } S < S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{634}$$
$$R_D = \begin{cases} P_{eq}, & \text{if } S = S_{max} \\ 0, & \text{otherwise} \end{cases} \tag{635}$$
$$E_t = min\left(max\left(0, E_{t,pot}\frac{S - sw}{sm - sw}\right), E_{t,pot}\right) \tag{636}$$
$$E_{t,pot} = E_p - E_i \tag{637}$$
$$L = K_{sat}S^c \tag{638}$$

Where S [mm] is the current storage in the soil moisture zone, refilled by infiltration F_i $[mm/d]$ and drained by Dunne-type runoff R_D $[mm/d]$, evapotranspiration E_t $[mm/d]$ and leakage L $[mm/d]$. F_i is calculated as equivalent precipitation P_{eq} minus Horton-type runoff R_H. P_{eq} is the sum of melt water M_w and the ground-heat flux G_s. R_H occurs at fixed rate ϕ $[mm/d]$ and only if the soil moisture is not saturated. R_D is equal to equivalent precipitation P_{eq} but occurs only when the store is at maximum capacity S_{max} [mm]. E_t fulfils any leftover evaporation demand after interception. E_t occurs at the potential rate until the plant stress point sm [mm], decreases linearly until the wilting point sw [mm] and is zero for any lower storage values. L has a non-linear relationship with storage through time parameter K_{sat} $[d^{-1}]$ and coefficient c [-].

$$\frac{S_{fast}}{dt} = L_f - R_f \tag{639}$$
$$L_f = L - L_s \tag{640}$$
$$L_s = min(L, L_{max}) \tag{641}$$
$$R_f = k_f * S_{fast} \tag{642}$$

Where S_{fast} [mm] is the current storage in the fast runoff reservoir, refilled by leakage-to-fast-flow L_f $[mm/d]$ and drained by fast runoff R_f $[mm/d]$. L_f depends on leakage L from soil moisture and the leakage-to-slow-flow L_s. L_s is calculated from a maximum leakage rate L_{max} $[mm/d]$. R_f has a linear relation with storage through time parameter k_f $[mm/d]$.

$$\frac{dS_{slow}}{dt} = L_s - R_s \tag{643}$$
$$R_s = k_s * S_{slow} \tag{644}$$

Where S_{slow} [mm] is the current storage in the slow runoff reservoir, refilled by leakage-to-slow-flow L_s $[mm/d]$ and drained by slow runoff R_s $[mm/d]$. R_s has a linear relation with storage through time parameter k_s $[mm/d]$. Total flow:

$$Q = R_H + R_D + R_F + R_S \tag{645}$$

S2.45 Precipitation-Runoff Modelling System (PRMS) (model ID: 45)

The PRMS model (fig. S46) is a modelling system that, in its most recent version, allows the user to specify a wide variety of catchment processes and flux equations (Markstrom et al., 2015). The version presented here is a simplified version of the original PRMS model (Leavesley et al., 1983). Simplifications involve the use of PET time series instead of within-model estimates based on temperature, and simpler interception and snow routines. The model has 7 stores and 18 parameters (TT, ddf, α, β, $STOR$, $RETIP$, SCN, SCX, $REMX$, $SMAX$, c_{gw}, $RESMAX$, k_1, k_2, k_3, k_4, k_5, k_6). The model aims to represent:

- Snow accumulation and melt;
- Interception by vegetation;
- Depression storage and impervious surface areas;
- Direct runoff based on catchment saturation;
- Infiltration into soil moisture and connection with deeper groundwater;
- Potentially non-linear interflow, baseflow and groundwater sink.

S2.45.1 File names

Model: m_45_prms_18p_7s
Parameter ranges: m_45_prms_18p_7s_parameter_ ranges

S2.45.2 Model equations

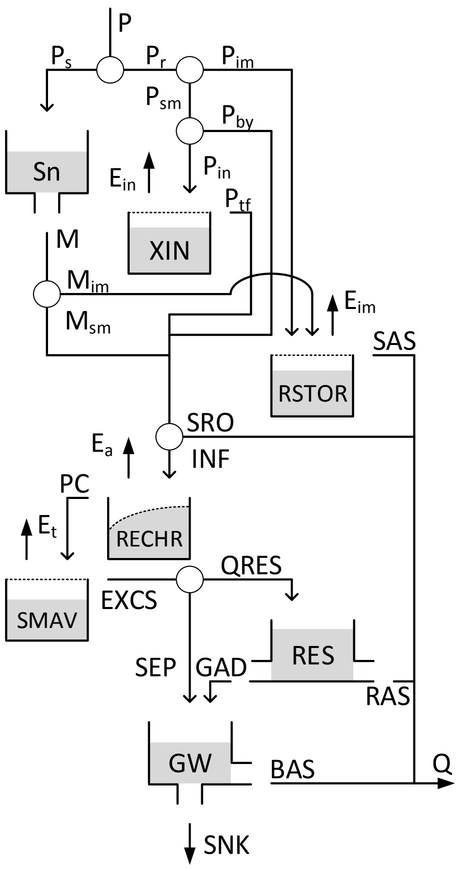

Figure S46: Structure of the PRMS model

$$\frac{dSn}{dt} = P_s - M \tag{646}$$

$$P_s = \begin{cases} P, & \text{if } T \leq TT \\ 0, & \text{otherwise} \end{cases} \tag{647}$$

$$M = \begin{cases} ddf * (T - TT), & \text{if } T \geq TT \\ 0, & \text{otherwise} \end{cases} \tag{648}$$

Where S is the current snow storage [mm], P_s the rain that falls as snow [mm], M the snowmelt [mm] based on a degree-day factor (ddf, [mm/°C/d]) and threshold temperature for snowfall and snowmelt (TT, [°C]).

$$\frac{dXIN}{dt} = P_{in} - E_{in} - P_{tf} \quad (649)$$

$$P_{in} = \alpha * P_{sm} \quad (650)$$

$$P_{sm} = \beta * P_r \quad (651)$$

$$P_r = \begin{cases} P, & \text{if } T > TT \\ 0, & \text{otherwise} \end{cases} \quad (652)$$

$$E_{in} = \begin{cases} \beta * E_p, & \text{if } XIN > 0 \\ 0, & \text{otherwise} \end{cases} \quad (653)$$

$$P_{tf} = \begin{cases} P_{in}, & \text{if } XIN = STOR \\ 0, & \text{otherwise} \end{cases} \quad (654)$$

Where XIN [mm] is the current storage in the interception reservoir, recharged by intercepted rainfall P_{in} $[mm/d]$ and drained by evaporation E_i $[mm/d]$ and throughfall P_{tf} $[mm/d]$. P_{in} $[mm/d]$ is the fraction α [-] of rainfall on non-impervious area P_{sm} $[mm/d]$ that does not bypass the interception reservoir. P_{sm} $[mm/d]$ is the fraction β [-] of rainfall P_r $[mm/d]$ that does not fall on impervious area. Rainfall is given as all precipitation P $[mm/d]$ that occurs when temperature T [°C] is above a threshold TT [°C]. E_i $[mm/d]$ occurs at the potential rate E_p, corrected for the fraction of the catchment where interception can occur. Throughfall P_{tf} is all rainfall that reaches the interception reservoir when it is at maximum capacity $STOR$ [mm].

$$\frac{dRSTOR}{dt} = P_{im} + M_{im} - E_{im} - SAS \quad (655)$$

$$P_{im} = (1 - \beta) * P_r \quad (656)$$

$$M_{im} = (1 - \beta) * M \quad (657)$$

$$E_{im} = \begin{cases} (1 - \beta) * E_p, & \text{if } RSTOR > 0 \\ 0, & \text{otherwise} \end{cases} \quad (658)$$

$$SAS = \begin{cases} P_{im} + M_{im}, & \text{if } RSTOR = RETIP \\ 0, & \text{otherwise} \end{cases} \quad (659)$$

Where $RSTOR$ [mm] is current depression storage, refilled by rainfall and snowmelt on impervious area, P_{im} $[mm/d]$ and M_{im} $[mm/d]$ respectively, and drained by evaporation E_{im} $[mm/d]$ and surface runoff SAS $[mm/d]$. P_{im} is given as the fraction $1-\beta$ of rainfall P_r. M_{im} is given as the fraction $1 - \beta$ of snowmelt M. E_{im} occurs at the potential rate E_p, corrected for the fraction of the catchment where impervious areas can occur. SAS occurs when the depression store is at maximum capacity $RETIP$ [mm].

$$\frac{dRECHR}{dt} = INF - E_a - PC \tag{660}$$

$$INF = M_{sm} + P_{tf} + P_{by} - SRO \tag{661}$$

$$M_{sm} = \beta * M \tag{662}$$

$$P_{by} = (1-\alpha) * P_{sm} \tag{663}$$

$$SRO = \left[SCN + (SCX - SCN) * \frac{RECHR}{REMX}\right] * (M_{sm} + P_{tf} + P_{by}) \tag{664}$$

$$E_a = \frac{RECHR}{REMX} * (E_p - E_i - E_{im}) \tag{665}$$

$$PC = \begin{cases} INF, & \text{if } RECHR = REMX \\ 0, & \text{otherwise} \end{cases} \tag{666}$$

Where $RECHR$ [mm] is the current storage in the upper soil moisture zone, recharged by infiltration INF $[mm/d]$ and drained by evaporation E_a $[mm/d]$ and percolation PC $[mm/d]$. INF is the difference between incoming snowmelt M_{sm} $[mm/d]$, throughfall P_{tf} $[mm/d]$ and interception bypass P_{by} $[mm/d]$, and surface runoff from saturated area SRO $[mm/d]$. S_{sm} is snowmelt from the fraction β [-] of the catchment that is not impervious. P_{by} is the fraction $1-\alpha$ of rainfall over non-impervious area P_{sm} that bypasses the interception store. SRO has a linear relation between minimum contributing area SCN [-] and maximum contributing area SCX [-] based on current storage $RECHR$ and maximum storage $REMX$ [mm]. E_a uses a similar linear relationship and accounts for already fulfilled evaporation demand by interception and impervious areas. PC occurs when the store reaches maximum capacity.

$$\frac{dSMAV}{dt} = PC - E_t - EXCS \tag{667}$$

$$E_t = \begin{cases} \frac{SMAV}{SMAX} * (E_p - E_{in} - E_{im} - E_a), & \text{if } RECHR < (E_p - E_{in} - E_{im}) \\ 0, & \text{otherwise} \end{cases} \tag{668}$$

$$EXCS = \begin{cases} PC, & \text{if } SMAV = SMAX - REMX \\ 0, & \text{otherwise} \end{cases} \tag{669}$$

Where $SMAV$ [mm] is the current storage in the lower soil moisture zone, recharged by percolation from the upper zone PC $[mm/d]$ and drained by transpiration E_t $[mm/d]$ and soil moisture excess $EXCS$ $[mm/d]$. E_t is corrected for already fulfilled evaporation demand and only occurs if the upper zone can not satisfy this demand. E_t uses a linear relationship between current storage and the maximum storage in the lower zone $SMAX - REMX$ [mm]. $EXCS$ only occurs when the store has reached maximum capacity $SMAX - REMX$.

$$\frac{dRES}{dt} = QRES - GAD - RAS \tag{670}$$
$$QRES = min(EXCS - SEP, 0) \tag{671}$$
$$GAD = k_1 \left(\frac{RES}{RESMAX}\right)^{k_2} \tag{672}$$
$$RAS = k_3 * RES + k_4 * RES^2 \tag{673}$$
$$\tag{674}$$

Where RES [mm] is the current storage in the runoff reservoir, filled by the difference between soil moisture excess $EXCS$ $[mm/d]$ and constant groundwater recharge SEP $[mm/d]$, and drained by groundwater drainage GAD $[mm/d]$ and interflow component RAS $[mm/d]$. GAD is potentially non-linear using time coefficient k_1 $[d^{-1}]$ and non-linearity coefficient k_2 [-], and is also scaled by the maximum reservoir capacity $RESMAX$ [mm]. RAS is non-linear interflow based on coefficients k_3 $[d^{-1}]$ and k_4 $[mm^{-1}d^{-1}]$.

$$\frac{dGW}{dt} = SEP + GAD - BAS - SNK \tag{675}$$
$$SEP = min(c_{gw}, EXCS) \tag{676}$$
$$BAS = k_5 * GW \tag{677}$$
$$SNK = k_6 * GW \tag{678}$$

Where GW [mm] is the current groundwater storage, refilled by groundwater recharge from soil moisture SEP and recharge from runoff reservoir GAD and drained by baseflow BAS $[mm/d]$ and flow to deeper groundwater SNK $[mm/d]$. SEP occurs at the maximum rate c_{gw} $[mm/d]$ if possible. BAS is a linear reservoir with time coefficient k_5 $[d^{-1}]$. SNK is a linear reservoir with time coefficient k_6 $[d^{-1}]$. Total flow Q_t $[mm/d]$:

$$Q_t = SAS + SRO + RAS + BAS \tag{679}$$

S2.46 Climate and Land-use Scenario Simulation in Catchments model (model ID: 46)

The CLASSIC model (fig. S47) is developed as a modular semi-distributed grid-based rainfall runoff model (Crooks and Naden, 2007). For comparability with other models the grid-based routing component is not included here, nor is the arable soil element because input data for this soil type is not supported. The model represents runoff from three different soil categories: permeable, semi-permeable and impermeable. It has 8 stores and 12 parameters (f_{ap}, f_{dp}, d_p, c_q, d_1, f_{as}, f_{ds}, d_s, d_2, c_{xq}, c_{xs}, c_u). The model aims to represent:

- Division into permeable, semi-permeable and impermeable areas;
- Infiltration into permeable soils and deficit-based soil moisture accounting;
- Infiltration into semi-permeable soils and direct runoff from semi-permeable soils (bypassing the moisture accounting);
- Fixed interception on impermeable soils;
- Linear flow routing from permeable soils;
- Fast and slow routing from semi-permeable soils;
- Linear flow routing from impermeable soils.

S2.46.1 File names

Model: m_46_classic_12p_8s
Parameter ranges: m_46_classic_12p_8s_parameter_ ranges

S2.46.2 Model equations

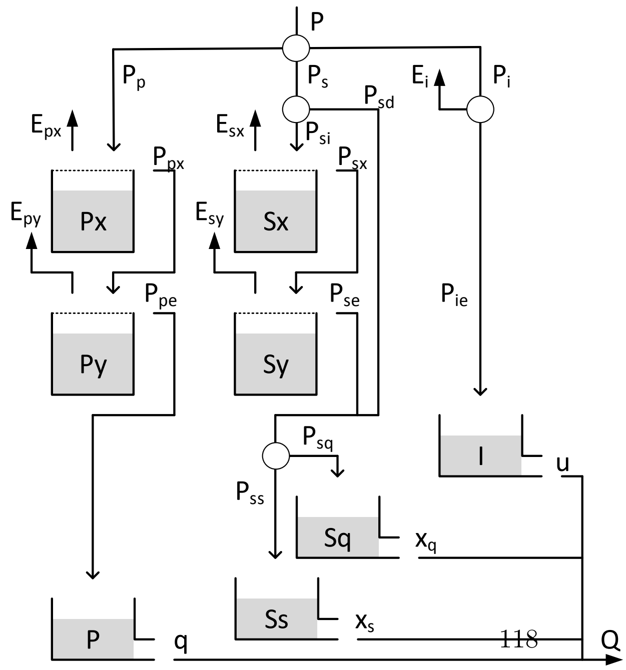

Figure S47: Structure of the CLASSIC model

$$\frac{dP_x}{dt} = P_p - E_{px} - P_{px} \quad (680)$$

$$P_p = f_{ap} * P \quad (681)$$

$$E_{px} = \begin{cases} f_{ap} * E_p, & \text{if } P_x > 0 \\ 0, & \text{otherwise} \end{cases} \quad (682)$$

$$P_{px} = \begin{cases} P_p, & \text{if } P_x = f_{dp} * d_p \\ 0, & \text{otherwise} \end{cases} \quad (683)$$

Where P_x [mm] is the current storage in the upper permeable layer, refilled by precipitation P_p $[mm/d]$ and drained by evaporation E_{px} $[mm/d]$ and excess flow P_{px} $[mm/d]$. P_p is the fraction of precipitation P $[mm/d]$ that falls on permeable area f_{ap} [-]. E_{px} occurs at the potential rate E_p $[mm/d]$ whenever possible, adjusted for the fraction of area that is permeable soil. P_{px} only occurs when the store is at maximum capacity $f_{dp} * d_p$, where d_p is the total soil depth (sum of depths X and Y) in the permeable area and f_{dp} the fraction of this depth that is store X.

$$\frac{dP_y}{dt} = -P_{px} + E_{py} + P_{pe} \tag{684}$$

$$E_{py} = 1.9 * exp\left[\frac{-0.6523 * (P_y + f_{dp} * d_p)}{f_{dp} * d_p}\right] * (f_{ap} * E_p - E_{px}) \tag{685}$$

$$P_{pe} = \begin{cases} P_{px}, & \text{if } P_y = 0 \\ 0, & \text{otherwise} \end{cases} \tag{686}$$

Where P_y [mm] is the current *deficit*, which is increased by evaporation E_{py} $[mm/d]$ and decreased by inflow P_{px} $[mm/d]$. Effective precipitation P_{pe} $[mm/d]$ is only generated when the deficit is 0. E_{py} decreases exponentially with increasing deficit.

$$\frac{dP}{dt} = P_{pe} - q \tag{687}$$

$$q = c_q * P \tag{688}$$

Where P [mm] is the current storage in the permeable soil routing store, refilled by effective rainfall on permeable soil P_{pe} $[mm/d]$ and drained by baseflow q $[mm/d]$. q has a linear relation with storage through time scale parameter c_p $[d^{-1}]$.

$$\frac{dS_x}{dt} = P_{si} - E_{sx} - P_{sx} \tag{689}$$

$$P_{si} = d_1 * P_s \tag{690}$$

$$P_s = f_{as} * P \tag{691}$$

$$E_{sx} = \begin{cases} f_{as} * E_p, & \text{if } S_x > 0 \\ 0, & \text{otherwise} \end{cases} \tag{692}$$

$$P_{sx} = \begin{cases} P_s, & \text{if } S_x = f_{ds} * d_s \\ 0, & \text{otherwise} \end{cases} \tag{693}$$

Where S_x [mm] is the current storage in the upper semi-permeable layer, refilled by infiltration P_{si} $[mm/d]$ and drained by evaporation E_{sx} $[mm/d]$ and excess flow P_{sx} $[mm/d]$. P_{si} is the fraction d_1 [mm] of precipitation on semi-permeable area P_s that infiltrates into the soil. The complementary fraction $1 - d_1$ of P_s bypasses the soil and directly becomes effective rainfall as P_{sd}. P_s is the fraction of precipitation P $[mm/d]$ that falls on semi-permeable area f_{as} [-] . E_{sx} occurs at the potential rate E_p $[mm/d]$ whenever possible, adjusted for the fraction of area that is semi-permeable soil. P_{sx} only occurs when the store is at maximum capacity $f_{ds} * d_s$, where d_s is the total soil depth (sum of depths X and Y) in the semi-permeable area and f_{ds} the fraction of this depth that is store X.

$$\frac{dS_y}{dt} = -P_{sx} + E_{sy} + P_{se} \tag{694}$$

$$E_{sy} = 1.9 * exp\left[\frac{-0.6523 * (S_y + f_{ds} * d_s)}{f_{ds} * d_s}\right] * (f_{as} * E_p - E_{sx}) \tag{695}$$

$$P_{pe} = \begin{cases} P_{sx}, & \text{if } S_y = 0 \\ 0, & \text{otherwise} \end{cases} \tag{696}$$

Where S_y [mm] is the current *deficit*, which is increased by evaporation E_{sy} $[mm/d]$ and decreased by inflow P_{sx} $[mm/d]$. Effective precipitation P_{se} $[mm/d]$ is only generated when the deficit is 0. E_{sy} decreases exponentially with increasing deficit.

$$\frac{dS_q}{dt} = P_{sq} - x_q \tag{697}$$

$$P_{sq} = d_2 * (P_{se} + P_{sd}) \tag{698}$$

$$x_q = c_{xq} * S_q \tag{699}$$

Where S_q [mm] is the current storage in the semi-permeable quick soil routing store, refilled by a fraction of effective rainfall on semi-permeable soil P_{sq} $[mm/d]$ and drained by quick flow x_q $[mm/d]$. P_{sq} is the fraction d_2 [-] of $(P_{se} + P_{Sd})$ that is quick flow. x_q has a linear relation with storage through time scale parameter c_{xq} $[d^{-1}]$.

$$\frac{dS_s}{dt} = P_{ss} - x_s \tag{700}$$

$$P_{ss} = (1 - d_2) * (P_{se} + P_{sd}) \tag{701}$$

$$x_s = c_{xs} * S_s \tag{702}$$

Where S_s [mm] is the current storage in the semi-permeable quick soil routing store, refilled by a fraction of effective rainfall on semi-permeable soil P_{ss} $[mm/d]$ and drained by slow flow x_s $[mm/d]$. P_{ss} is the fraction $1 - d_2$ [-] of $(P_{se} + P_{Sd})$ that is slow flow. x_s has a linear relation with storage through time scale parameter c_{xs} $[d^{-1}]$.

$$\frac{dI}{dt} = P_{ie} - u \tag{703}$$

$$P_{ie} = P_i - E_i \tag{704}$$

$$P_i = P - P_p - P_s \tag{705}$$

$$u = c_u * I \tag{706}$$

Where I [mm] is the current storage in the impermeable soil routing store, refilled by effective rainfall on impermeable soil P_{ie} $[mm/d]$ and drained by baseflow u $[mm/d]$. P_{ie} is the remained of precipitation on impermeable soils P_i $[mm/d]$, after a constant evaporation E_i has been extracted. E_i is fixed at 0.5 $[mm/d]$. x_s has a linear relation with storage through time scale parameter c_{xs} $[d^{-1}]$. Total flow:

$$Q = q + x_s + x_q + u \tag{707}$$

S3 Flux equations

Section S2 gives descriptions of each model and provides both Ordinary Differential Equations and the constitutive functions that describe each model's fluxes. These constitutive functions and any relevant constraints are implemented in MARRMoT as individual *flux files*. Each *flux file* contains computer code that combines the consitituve function and constraints (if needed). *Flux files* are located in the folder "./MARRMoT/Models/Flux files/". The User Manual contains details on understanding, modifying and creating new *flux files*. Table S1 shows a complete overview of fluxes currently implemented in MARRMoT.

Table S1: Equations from model descriptions and their implementation in MARRMoT (Table starts on following page)

Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
Abstraction	Groundwater abstraction at a constant rate	abstraction_1	$flux_{out} = \theta_1$	None, taken from a store with possible negative depth	$flux_{out} = \theta_1$	25
Baseflow	Linear reservoir	baseflow_1	$flux_{out} = \theta_1 * S$		$flux_{out} = \theta_1 * S$	2, 4, 6, 8, 9, 12, 13, 15, 16, 17, 18, 20, 21, 24, 25, 26, 27, 28, 29, 30, 31, 32, 33, 34, 35, 36, 37, 38, 40, 41, 43, 44, 45, 46
	Non-linear outflow from a reservoir	baseflow_2	$flux_{out} = \left(\frac{1}{\theta_1}S\right)^{\frac{1}{\theta_2}}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\frac{S}{\Delta t}, \left(\frac{1}{\theta_1}max(S,0)\right)^{\frac{1}{\theta_2}}\right)$	9, 11
	Empirical exponential outflow from a reservoir	baseflow_3	$flux_{out} = \frac{S_{max}^{-4}}{4}S^5$	Empirical equation, so interwoven with other equations that no constraints are needed. Also implicitly assumes time step $\Delta t = 1$	$flux_{out} = \frac{S_{max}^{-4}}{4}S^5$	7
	Exponential outflow from a deficit store	baseflow_4	$flux_{out} = \theta_1 e^{-\theta_2 S}$		$flux_{out} = \theta_1 e^{-\theta_2 S}$	14
	Non-linear outflow scaled by current relative storage	baseflow_5	$flux_{out} = \theta_1\left(\frac{S}{S_{max}}\right)^{\theta_2}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1\left(\frac{max(0,S)}{S_{max}}\right)^{\theta_1}\right)$	22
	Quadratic outflow from reservoir if a storage threshold is exceeded	baseflow_6	$flux_{out} = \begin{cases}\theta_1 * S^2, & if\ S > \theta_2 \\ 0, & otherwise\end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = \min\left(\theta_1 * S^2, \frac{S}{\Delta t}\right) * [1 - logisticSmoother\_S(S,\theta_2)]$	25
	Non-linear outflow from a reservoir	baseflow_7	$flux_{out} = \theta_1 S^{\theta_2}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1 max(0,S)^{\theta_2}\right)$	39, 42

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Exponential scaled outflow from a deficit store	baseflow_8	$flux_{out} = \theta_1 \left(e^{\theta_2 S / S_{max}} - 1\right)$	$S \leq S_{max}$ $S \geq 0$	$flux_{out} = \theta_1 \left(e^{\theta_2 * min\left(1, max\left(0, S / S_{max}\right)\right)} - 1\right)$	23
	Linear outflow from a reservoir if a storage threshold is exceeded	baseflow_9	$flux_{out} = \begin{cases} \theta_1 (S - \theta_2), & if\ S > \theta_2 \\ 0, & otherwise \end{cases}$		$flux_{out} = \theta_1 * max(0, S - \theta_2)$	20
Capillary rise	Capillary rise scaled by relative deficit in receiving store	capillary_1	$flux_{out} = \theta_1 \left[1 - \frac{S_1}{S_{1,max}}\right]$	$flux_{out} \leq \frac{S_2}{\Delta t}$	$flux_{out} = min\left(\theta_1 \left[1 - \frac{S_1}{S_{1,max}}\right], \frac{S_2}{\Delta t}\right)$	37
	Capillary rise at a constant rate	capillary_2	$flux_{out} = \begin{cases} \theta_1, & if\ S \geq 0 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1\right)$	13, 15
	Capillary rise if the receiving store is below a storage threshold	capillary_3	$flux_{out} = \begin{cases} \theta_1 \left(1 - \frac{S_1}{\theta_2}\right), & if\ S_1 < \theta \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S_2}{\Delta t}$	$flux_{out} = min\left(\frac{S_2}{\Delta t}, \theta_1 \left(1 - \frac{S_1}{\theta_2}\right) * logisticSmoother\_S(S_1, \theta_2)\right)$	38
Depression storage	Exponential inflow rate into surface depressions	depression_1	$flux_{out} = \theta_1 * exp\left[-\theta_2 \frac{S}{S_{max} - S}\right] * flux_{in}$	$flux_{out} \leq \frac{S_{max} - S}{\Delta t}$ $S \leq S_{max}$	$flux_{out} = min\left(\theta_1 * exp\left[-\theta_2 \frac{S}{\max(S_{max} - S, 0)}\right] * flux_{in}, \frac{S_{max} - S}{\Delta t}\right)$	36
Evaporation	Evaporation at the potential rate	evap_1	$E_a = \begin{cases} E_p, & if\ S \geq 0 \\ 0, & otherwise \end{cases}$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(E_p, \frac{S}{\Delta t}\right)$	2, 6, 12, 13, 16, 17, 18, 23, 25, 26, 27, 33, 34, 36, 38, 39, 41, 42, 44, 45, 46
	Evaporation at scaled plant-controlled rate	evap_2	$E_a = \theta_1 \frac{S}{S_{max}}$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1 \frac{S}{S_{max}}, E_p, \frac{S}{\Delta t}\right)$	18, 36
	Evaporation scaled by relative storage below a wilting point and at the potential rate above wilting point	evap_3	$E_A = \begin{cases} E_p \frac{S}{\theta_1 S_{max}}, & if\ S < \theta_1 S_{max} \\ E_p, & otherwise \end{cases}$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(E_p \frac{S}{\theta_1 S_{max}}, E_p, \frac{S}{\Delta t}\right)$	3, 11, 14, 21, 26, 34, 37, 42
	Scaled evaporation if storage is above the wilting point, constrained by a limitation parameter	evap_4	$E_a = E_p * max\left(0, \theta_1 \frac{S - \theta_2 S_{max}}{S_{max} - \theta_2 S_{max}}\right)$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(E_p * max\left(0, \theta_1 \frac{S - \theta_2 S_{max}}{S_{max} - \theta_2 S_{max}}\right), \frac{S}{\Delta t}\right)$	15
	Evaporation from bare soil, scaled by relative storage	evap_5	$E_a = (1 - \theta_1) \frac{S}{S_{max}} E_p$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left((1 - \theta_1) \frac{S}{S_{max}} E_p, \frac{S}{\Delta t}\right)$	4, 8, 9, 16

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Transpiration from vegetation at the potential rate if storage is above a wilting point and scaled by relative storage if not	evap_6	$E_A = \begin{cases} \theta_1 * E_p, & if\ S > \theta_2 * S_{max} \\ \theta_1 \frac{S}{\theta_2 S_{max}} E_p, & otherwise \end{cases}$	$E_a \leq \theta_1 E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1 E_p \frac{S}{\theta_2 S_{max}}, \theta_1 E_p, \frac{S}{\Delta t}\right)$	4, 9, 16
	Evaporation scaled by relative storage	evap_7	$E_a = \frac{S}{S_{max}} E_p$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\frac{S}{S_{max}} E_p, \frac{S}{\Delta t}\right)$	1, 3, 10, 11, 19, 22, 24, 29, 30, 31, 32, 33, 35, 45
	Transpiration from vegetation, at potential rate if soil moisture is above the wilting point, and linearly decreasing if not. Also scaled by relative storage across all stores	evap_8	$E_A = \begin{cases} \frac{S_1}{S_1 + S_2} \theta_1 E_p, & if\ S_1 > \theta_2 \\ \frac{S_1}{\theta_2} * \frac{S_1}{S_1 + S_2} \theta_1 E_p, & otherwise \end{cases}$	$E_a \leq \frac{S_1}{\Delta t}$ $E_a \geq 0$	$E_a = max\left(min\left(\frac{S_1}{S_1 + S_2} \theta_1 E_p, \frac{S_1}{\theta_2} * \frac{S_1}{S_1 + S_2} \theta_1 E_p, \frac{S_1}{\Delta t}\right), 0\right)$	8
	Evaporation from bare soil scaled by relative storage and by relative water availability across all stores	evap_9	$E_a = \frac{S_1}{S_1 + S_2} * (1 - \theta_1) \frac{S_1}{S_{max} - S_2} E_p$	$E_a \leq \frac{S_1}{\Delta t}$ $E_a \geq 0$	$E_a = max\left(min\left(\frac{S_1}{S_1 + S_2} * (1 - \theta_1) \frac{S_1}{S_{max} - S_2} E_p, \frac{S_1}{\Delta t}\right), 0\right)$	8
	Evaporation from bare soil, scaled by relative storage	evap_10	$E_a = \theta_1 \frac{S}{S_{max}} E_p$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1 \frac{S}{S_{max}} E_p, \frac{S}{\Delta t}\right)$	8
	Evaporation quadratically related to current soil moisture	evap_11	$E_a = \left(2 \frac{S}{S_{max}} - \left(\frac{S}{S_{max}}\right)^2\right) E_p$	$E_a \geq 0$	$E_a = max\left(0, \left(2 \frac{S}{S_{max}} - \left(\frac{S}{S_{max}}\right)^2\right) E_p\right)$	7
	Evaporation from deficit store, with exponential decline as deficit goes below a threshold	evap_12	$E_a = min\left(1, e^{2\left(1 - \frac{S}{\theta_1}\right)}\right) E_p$		$E_a = min\left(1, e^{2\left(1 - \frac{S}{\theta_1}\right)}\right) E_p$	5
	Exponentially scaled evaporation	evap_13	$E_a = \theta_1^{\theta_2} E_p$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1^{\theta_2} E_p, \frac{S}{\Delta t}\right)$	40
	Exponentially scaled evaporation that only activates if another store goes below a certain threshold	evap_14	$E_A = \begin{cases} \theta_1^{\theta_2} E_p, & if\ S_2 \leq S_{2,min} \\ 0, & otherwise \end{cases}$	$E_a \leq \frac{S_1}{\Delta t}$	$E_a = min\left(\theta_1^{\theta_2} E_p, \frac{S_1}{\Delta t}\right) * logisticSmoother\_S(S_2, S_{2,min})$	40
	Scaled evaporation if another store is below a threshold	evap_15	$E_a = \begin{cases} \frac{S_1}{S_{max}} E_p, & if\ S_2 < \theta_1 \\ 0, & otherwise \end{cases}$	$E_a \leq \frac{S_1}{\Delta t}$	$E_a = min\left(\frac{S_1}{S_{1,max}} * E_p * logisticSmoother\_S(S_2, \theta_2), \frac{S_1}{\Delta t}\right)$	41, 45

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Scaled evaporation if another store is below a threshold	evap_16	$E_a = \begin{cases} \theta_1 E_p, & if\ S_2 < \theta_2 \\ 0, & otherwise \end{cases}$	$E_a \leq \frac{S_1}{\Delta t}$	$E_a = min\left(\theta_1 * E_p * logisticSmoother\_S(S_2, \theta_2), \frac{S_1}{\Delta t}\right)$	17, 25
	Scaled evaporation from a store that allows negative values	evap_17	$E_a = \frac{1}{1 + e^{-\theta_1 * S}} E_p$	None, because the store is allowed to go negative	$E_a = \frac{1}{1 + e^{-\theta_1 * S}} E_p$	39
	Exponentially declining evaporation from deficit store	evap_18	$E_a = \theta_1 e^{\frac{-\theta_2 S}{\theta_3}} E_p$		$E_a = \theta_1 e^{\frac{-\theta_2 S}{\theta_3}} E_p$	46
	Non-linear scaled evaporation	evap_19	$E_a = \theta_1 \left(\frac{S}{S_{max}}\right)^{\theta_2} E_p$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1 * max\left(0, \frac{S}{S_{max}}\right)^{\theta_2} E_p, E_p, \frac{S}{\Delta t}\right)$	23, 43
	Evaporation limited by a maximum evaporation rate and scaled below a wilting point	evap_20	$E_A = \begin{cases} \theta_1 \frac{S}{\theta_2 S_{max}}, & if\ S < \theta_2 S_{max} \\ E_p, & otherwise \end{cases}$	$E_a \leq E_p$ $E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\theta_1 \frac{S}{\theta_2 S_{max}}, E_p, \frac{S}{\Delta t}\right)$	20
	Threshold-based evaporation with constant minimum rate	evap_21	$E_a = \begin{cases} E_p, & if\ S > \theta_1 \\ \frac{S}{\theta_1} E_p, & if\ \theta_2 \theta_1 \geq S \geq \theta_1 \\ \theta_2\ E_p & otherwise \end{cases}$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(max\left(\theta_2, min\left(\frac{S}{\theta_1}, 1\right)\right) * E_p, \frac{S}{\Delta t}\right)$	28
	Threshold-based evaporation rate	evap_22	$E_a = \begin{cases} E_p, & if\ S > \theta_1 \\ \frac{S - \theta_1}{\theta_1 - \theta_2} E_p, & if\ \theta_2 \theta_1 \geq S \geq \theta_1 \\ 0 & otherwise \end{cases}$	$E_a \leq \frac{S}{\Delta t}$	$E_a = min\left(\frac{S}{\Delta t}, min\left(E_p, max\left(0, \frac{S - \theta_1}{\theta_2 - \theta_1} E_p\right)\right)\right)$	44
Exchange	Water exchange between aquifer and channel	exchange_1	$flux_{out} = \begin{cases} \theta_1 * \left\lvert\frac{S}{\Delta t}\right\rvert + \theta_2 \left(1 - exp\left[-\theta_3 * \left\lvert\frac{S}{\Delta t}\right\rvert\right]\right), & if\ S \geq 0 \\ -\left[\theta_1 * \left\lvert\frac{S}{\Delta t}\right\rvert + \theta_2 \left(1 - exp\left[-\theta_3 * \left\lvert\frac{S}{\Delta t}\right\rvert\right]\right)\right], & if\ S < 0 \end{cases}$	$\begin{cases} No\ constraint \\ flux_{out} \leq flux_{in} \end{cases}$ The “channel” store in this model has 0 time delay, so the incoming flux to the channel is the maximum channel-to-groundwater flux size. Groundwater has infinite depth	$flux_{out} = \max\left(\left[\theta_1 * \left\lvert\frac{S}{\Delta t}\right\rvert + \theta_2 * \left(1 - exp\left[-\theta_3 * \left\lvert\frac{S}{\Delta t}\right\rvert\right]\right)\right] * sign(S), -flux_{in}\right)$	36
	Water exchange based on relative storages	exchange_2	$flux_{out} = \theta_1 \left(\frac{S_1}{S_{1,max}} - \frac{S_2}{S_{2,max}}\right)$		$flux_{out} = \theta_1 \left(\frac{S_1}{S_{1,max}} - \frac{S_2}{S_{2,max}}\right)$	38
	Water exchange with infinite size store based on threshold	exchange_3	$flux_{out} = \theta_1 * (S - \theta_2)$		$flux_{out} = \theta_1 * (S - \theta_2)$	36

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
Infiltration	Infiltration as exponentially declining based on relative storage (taken from a flux)	infiltration_1	$flux_{out} = \theta_1 * exp\left[-\theta_2 \frac{S}{S_{max}}\right]$	$flux_{out} \leq flux_{in}$	$flux_{out} = min\left(\theta_1 * exp\left[-\theta_2 \frac{S}{S_{max}}\right], flux_{in}\right)$	18, 36, 44
	Delayed infiltration as exponentially declining based on relative storage (taken from a store)	infiltration_2	$flux_{out} = \theta_1 * exp\left[-\theta_2 \frac{S_1}{S_{1,max}}\right] - flux_{used}$	$0 \leq flux_{out} \leq \frac{S_2}{\Delta t}$	$flux_{out} = \max\left(min\left(\theta_1 * exp\left[-\theta_2 \frac{S_1}{S_{1,max}}\right] - flux_{used}, \frac{S_2}{\Delta t}\right), 0\right)$	36
	Infiltration to soil moisture of liquid water stored in snow pack	infiltration_3	$flux_{out} = \begin{cases} flux_{in}, & if\ S \geq S_{max} \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in}[1 - logisticSmoother\_S(S, S_{\max})]$	37
	Constant infiltration rate	infiltration_4	$flux_{out} = \theta_1$	$flux_{out} \leq flux_{in}$	$flux_{out} = min(flux_{in}, \theta_1)$	15, 23, 40, 44
	Maximum infiltration rate non-linearly based on relative deficit and storage	infiltration_5	$flux_{out} = \theta_1\left(1 - \frac{S_1}{S_{1,max}}\right)\left(\frac{S_2}{S_{2,max}}\right)^{-\theta_2}$	To prevent complex numbers, S = [0,∞> To prevent numerical issues with a theoretical infinite infiltration rate, $flux_{out}$ < 10^9	$flux_{out} = min\left(10^9, \theta_1\left(1 - \frac{S_1}{S_{1,max}}\right) max\left(0, \frac{S_2}{S_{2,max}}\right)^{-\theta_2}\right)$	23
	Infiltration rate non-linearly scaled by relative storage	infiltration_6	$flux_{out} = \theta_1\left(\frac{S}{S_{max}}\right)^{\theta_2} flux_{in}$	$flux_{out} \leq flux_{in}$	$flux_{out} = min\left(\theta_1 * max\left(0, \frac{S}{S_{max}}\right)^{\theta_2} flux_{in}, flux_{in}\right)$	43
Interception	Interception excess when maximum capacity is reached	interception_1	$flux_{out} = \begin{cases} flux_{in}, & if\ S \geq S_{max} \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in}[1 - logisticSmoother\_S(S, S_{\max})]$	16, 18, 22, 26, 34, 36, 39, 42, 44, 45
	Interception excess after a constant amount is intercepted	interception_2	$flux_{out} = \begin{cases} flux_{in} - \theta_1, & if\ flux_{in} \geq 0 \\ 0, & otherwise \end{cases}$	$flux_{out} \geq 0$	$flux_{out} = max(flux_{in} - \theta_1, 0)$	2, 13, 15
	Interception excess after a fraction is intercepted	interception_3	$flux_{out} = \theta_1$		$flux_{out} = \theta_1$	8
	Interception excess after a time-varying fraction is intercepted	interception_4	$flux_{out} = \left(\theta_1 + (1 - \theta_1) * \cos\left(2\pi \frac{t * \Delta t - \theta_2}{t_{max}}\right)\right) * flux_{in}$	$flux_{out} \geq 0$	$flux_{out} = max\left(0, \theta_1 + (1 - \theta_1) * \cos\left(2\pi \frac{t * \Delta t - \theta_2}{t_{max}}\right)\right) * flux_{in}$	32, 35
	Interception excess after a combined absolute amount and fraction are intercepted	interception_5	$flux_{out} = \begin{cases} \theta_1 * flux_{in} - \theta_2, & if\ flux_{in} \geq 0 \\ 0, & otherwise \end{cases}$	$flux_{out} \geq 0$	$flux_{out} = max(\theta_1 * flux_{in} - \theta_2, 0)$	23

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
Interflow	Interflow as a scaled fraction of an incoming flux	interflow_1	$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$		$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$	18, 36
	Non-linear interflow	interflow_2	$flux_{out} = \theta_1 S^{(1+\theta_2)}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\theta_1 max(S,0)^{(1+\theta_2)}, max\left(\frac{S}{\Delta t},0\right)\right)$	37
	Non-linear interflow (variant)	interflow_3	$flux_{out} = \theta_1 S^{\theta_2}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\theta_1 max(S,0)^{\theta_2}, max\left(\frac{S}{\Delta t},0\right)\right)$	10, 19, 42, 43
	Combined linear and scaled quadratic interflow	interflow_4	$flux_{out} = \theta_1 S + \theta_2 S^2$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\theta_1 max(S,0) + \theta_2 max(S,0)^2, max\left(\frac{S}{\Delta t},0\right)\right)$	45
	Linear interflow	interflow_5	$flux_{out} = \theta_1 * S$		$flux_{out} = \theta_1 * S$	28, 33, 41
	Scaled linear interflow if a storage in the receiving store exceeds a threshold	interflow_6	$flux_{out} = \begin{cases} \theta_1 * S_1 * \frac{S_2/S_{2,max} - \theta_2}{1-\theta_2}, & if\ S_2/S_{2,max} > \theta_2 \\ 0, & otherwise \end{cases}$	$\frac{S_2}{S_{2,max}} \leq 1$	$flux_{out} = \left(\theta_1 * S_1 * \frac{min\left(1, S_2/S_{2,max}\right) - \theta_2}{1-\theta_2}\right) * \left[1 - logisticSmoother\_S\left(\frac{S_2}{S_{2,max}}, \theta_2\right)\right]$	41
	Non-linear interflow if storage exceeds a threshold	interflow_7	$flux_{out} = \begin{cases} \left(\frac{S - \theta_1 S_{max}}{\theta_2}\right)^{\frac{1}{\theta_3}}, & if\ S > \theta_1 S_{max} \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S - \theta_1 S_{max}}{\Delta t}$ To prevent complex numbers, S-θ1Smax = [0,∞>	$flux_{out} = min\left(max\left(0, \frac{S - \theta_1 S_{max}}{\Delta t}\right), \left(\frac{max(0, S - \theta_1 S_{max})}{\theta_2}\right)^{\frac{1}{\theta_3}}\right)$	9
	Linear interflow if storage exceeds a threshold	interflow_8	$flux_{out} = \begin{cases} \theta_1(S - \theta_2), & if\ S > \theta_2 \\ 0, & otherwise \end{cases}$		$flux_{out} = max\left(0, \theta_1(S - \theta_2)\right)$	3, 12, 27, 38
	Non-linear interflow if storage exceeds a threshold (variant)	interflow_9	$flux_{out} = \begin{cases} \left(\theta_1(S - \theta_2)\right)^{\theta_3}, & if\ S > \theta_2 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S - \theta_2}{\Delta t}$ To prevent complex numbers, S-θ2 = [0,∞>	$flux_{out} = min\left(\frac{S-\theta_2}{\Delta t}, \left(\theta_1 * max(0, S - \theta_2)\right)^{\theta_3}\right)$	4, 11, 16, 39
	Scaled linear interflow if storage exceeds a threshold	interflow_10	$flux_{out} = \begin{cases} \theta_1 \frac{(S - \theta_2)}{\theta_3}, & if\ S > \theta_2 \\ 0, & otherwise \end{cases}$		$flux_{out} = \theta_1 \frac{max(0, S - \theta_2)}{\theta_3}$	14
	Constant interflow if storage exceeds a threshold	interflow_11	$flux_{out} = \begin{cases} \theta_1, & if\ S > \theta_2 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S - \theta_2}{\Delta t}$	$flux_{out} = \min\left(\theta_1, \frac{S - \theta_2}{\Delta t}\right) * [1 - logisticSmoother\_S(S, \theta_2)]$	20

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
Misc	Auxiliary function to find contributing area	area_1	$A = \begin{cases} \theta_1 \left[\frac{S - S_{min}}{S_{max} - S_{min}}\right]^{\theta_2}, & if\ S > S_{min} \\ 0, & otherwise \end{cases}$	$A \leq 1$	$A = min\left(1, \theta_1 \left[\frac{S - S_{min}}{S_{max} - S_{min}}\right]^{\theta_2}\right) * [1 - logisticSmoother\_S(S, S_{min}]$	23
	General effective flow (returns flux [mm/d])	effective_1	$flux_{out} = \begin{cases} flux_{in,1} - flux_{in,2}, & if\ flux_{in,1} > flux_{in,2} \\ 0, & otherwise \end{cases}$		$flux_{out} = max(0, flux_{in,1} - flux_{in,2})$	22, 23, 25, 39, 40, 42, 43, 44, 45, 46
	Storage excess when store size changes (returns flux [mm/d])	excess_1	$flux_{out} = \frac{S - S_{max,new}}{\Delta t}$	$flux_{out} \geq 0$	$flux_{out} = max\left(\frac{S - S_{max,new}}{\Delta t}, 0\right)$	10, 19, 22, 37, 44
	Phenology-based correction factor for potential evapotranspiration (returns flux [mm/d])	phenology_1	$E_p^* = \begin{cases} 0, & if\ T(t) < \theta_1 \\ \frac{T(t) - \theta_1}{\theta_2 - \theta_1} * E_p, & if\ \theta_1 \leq T(t) < \theta_2 \\ E_p, & if\ T(t) \geq \theta_2 \end{cases}$		$E_p^* = min\left(1, max\left(0, \frac{T(t) - \theta_1}{\theta_2 - \theta_1}\right)\right) * E_p$	35
	Phenology-based maximum interception capacity (returns store size [mm])	phenology_2	$S_{max} = \theta_1 \left(1 + \theta_2 sin\left(2\pi \frac{t * \Delta t - \theta_3}{t_{max}}\right)\right)$	Assumes $0 \leq \theta_2 \leq 1$ to guarantee $S_{max} \geq 0$	$S_{max} = \theta_1 \left(1 + \theta_2 sin\left(2\pi \frac{t * \Delta t - \theta_3}{t_{max}}\right)\right)$	22
	Split flow (returns flux [mm/d])	split_1	$flux_{out} = \theta_1 * flux_{in}$		$flux_{out} = \theta_1 * flux_{in}$	5, 11, 13, 17, 21, 25, 26, 28, 29, 33, 34, 40, 41, 42, 43, 45, 46
Percolation	Percolation at a constant rate	percolation_1	$flux_{out} = \begin{cases} \theta_1, & if\ S \geq 0 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1\right)$	37
	Percolation scaled by current relative storage	percolation_2	$flux_{out} = \theta_1 \frac{S}{S_{max}}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1 \frac{S}{S_{max}}\right)$	21, 26, 34
	Non-linear percolation (empirical)	percolation_3	$flux_{out} = \frac{S_{max}^{-4}}{4}\left(\frac{4}{9}\right)^{-4} S^5$		$flux_{out} = \frac{S_{max}^{-4}}{4}\left(\frac{4}{9}\right)^{-4} S^5$	7
	Demand-based percolation scaled by available moisture	percolation_4	$flux_{out} = \frac{S}{S_{max}}\left[\theta_1 \left\{1 + \theta_2 \left(\frac{\sum deficiencies}{\sum capacities}\right)^{\theta_3}\right\}\right]$	$flux_{out} \leq \frac{S}{\Delta t}$ To avoid erratic numerical behaviour, $flux_{out} \geq 0$	$flux_{out} = max\left(0, min\left(\frac{S}{\Delta t}, \frac{max(S,0)}{S_{max}} * \left[\theta_1 \left\{1 + \theta_2 \left(\frac{\sum deficiencies}{\sum capacities}\right)^{\theta_3}\right\}\right]\right)\right)$	33
	Non-linear percolation	percolation_5	$flux_{out} = \theta_1 \left(\frac{S}{S_{max}}\right)^{\theta_2}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1 \left(\frac{max(0,S)}{S_{max}}\right)^{\theta_2}\right)$	22

continued …

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Threshold-based percolation from a store that can reach negative values	percolation_6	$flux_{out} = \begin{cases} \theta_1, & if\ S \geq \theta_2 \\ \theta_1 \frac{S}{\theta_2}, & if\ 0 < S < \theta_2 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1 min\left[1, \frac{max(0,S)}{\theta_2}\right]\right)$	39
Recharge	Recharge as scaled fraction of incoming flux	recharge_1	$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$		$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$	18, 36
	Recharge as non-linear scaling of incoming flux	recharge_2	$flux_{out} = \left(\frac{S}{S_{max}}\right)^{\theta_1} * flux_{in}$	To prevent complex numbers, S = [0,∞>	$flux_{out} = \left(\frac{max(0,S)}{S_{max}}\right)^{\theta_1} * flux_{in}$	7, 37, 45
	Linear recharge	recharge_3	$flux_{out} = \theta_1 * S$		$flux_{out} = \theta_1 * S$	19, 23, 24, 27, 30, 31, 32, 35, 38, 42
	Constant recharge from a store	recharge_4	$flux_{out} = \begin{cases} \theta_1, & if\ S \geq 0 \\ 0, & otherwise \end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1\right)$	23, 44
	Recharge to fulfil evaporation demand if the receiving store is below a threshold	recharge_5	$flux_{out} = \begin{cases} \theta_1 S_1 \left(1 - \frac{S_2}{\theta_2}\right), & if\ S_2 < \theta_2 \\ 0, & otherwise \end{cases}$		$flux_{out} = \theta_1 S_1 \left[1 - min\left(1, \frac{S_2}{\theta_2}\right)\right]$	20
	Non-linear recharge	recharge_6	$flux_{out} = \theta_1 S^{\theta_2}$	$flux_{out} \leq \frac{S}{\Delta t}$ To prevent complex numbers, S = [0,∞>	$flux_{out} = min\left(\theta_1 max(S,0)^{\theta_2}, max\left(\frac{S}{\Delta t}, 0\right)\right)$	44
	Constant recharge from a flux	recharge_7	$flux_{out} = \theta_1$	$flux_{out} \leq flux_{in}$	$flux_{out} = min(flux_{in}, \theta_1)$	45
Routing	Threshold-based non-linear routing	routing_1	$flux_{out} = \begin{cases} \theta_1 S^{\theta_2}, & if\ flux_{out} < \theta_3 S \\ \theta_3 S, & otherwise \end{cases}$	$flux_{out} \leq \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1 max(S,0)^{\theta_2}, \theta_3 \frac{S}{\Delta t}\right)$	39
Saturation excess	Saturation excess from a store that has reached maximum capacity	saturation_1	$flux_{out} = \begin{cases} flux_{in}, & if\ S \geq S_{max} \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in}[1 - logisticSmoother\_S(S, S_{max})]$	1, 3, 4, 6, 8, 9, 10, 11, 12, 14, 15, 16, 17, 18, 19, 20, 22, 24, 25, 30, 31, 32, 33, 35, 36, 39, 40, 41, 44, 45, 46

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Saturation excess from a store with different degrees of saturation	saturation_2	$flux_{out} = \left(1-\left(1-\frac{S}{S_{max}}\right)^{\theta_1}\right) * flux_{in}$	To prevent complex numbers, S/Smax = [0,∞>	$flux_{out} = \left(1-\left(min\left(1, max\left(0, \left(1-\frac{S}{S_{max}}\right)\right)\right)\right)^{\theta_1}\right) * flux_{in}$	2, 13, 22, 28, 29
	Saturation excess from a store with different degrees of saturation (exponential variant)	saturation_3	$flux_{out} = \left(1-\frac{1}{1+exp\left(\frac{S/S_{max}+0.5}{\theta_1}\right)}\right) * flux_{in}$		$flux_{out} = \left(1-\frac{1}{1+exp\left(\frac{S/S_{max}+0.5}{\theta_1}\right)}\right) * flux_{in}$	21, 26, 34
	Saturation excess from a store with different degrees of saturation (quadratic variant)	saturation_4	$flux_{out} = \left(1-\left(\frac{S}{S_{max}}\right)^2\right) * flux_{in}$	$0 \leq flux_{out}$	$flux_{out} = max\left(0, \left(1-\left(\frac{S}{S_{max}}\right)^2\right) * flux_{in}\right)$	7
	Deficit store: exponential saturation excess based on current storage and a threshold parameter	saturation_5	$flux_{out} = \left(1-min\left(1, \left(\frac{S}{\theta_1}\right)^{\theta_2}\right)\right) * flux_{in}$	To prevent complex numbers, S = [0,∞>	$flux_{out} = \left(1-min\left(1, \left(\frac{max(S,0)}{\theta_1}\right)^{\theta_2}\right)\right) * flux_{in}$	5
	Saturation excess from a store with different degrees of saturation (linear variant)	saturation_6	$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$		$flux_{out} = \theta_1 \frac{S}{S_{max}} * flux_{in}$	40
	Saturation excess from a store with different degrees of saturation (gamma function variant)	saturation_7	$flux_{out} = flux_{in} \begin{cases} \int_{x=\theta_5 * S+\theta_4}^{x=\infty} \frac{1}{\theta_1 \Gamma(\theta_2)} \left(\frac{x-\theta_3}{\theta_1}\right)^{\theta_2-1} e^{\left(-\frac{x-\theta_3}{\theta_1}\right)}, & x > \theta_3 \\ 0 \end{cases}$	To prevent numerical problems, S = [0,∞>	$flux_{out} = flux_{in} * integral\left(\frac{1}{\theta_1 \Gamma(\theta_2)} \left(\frac{max(x-\theta_3,0)}{\theta_1}\right)^{\theta_2-1} * e^{\left(-1*\frac{max(x-\theta_3,0)}{\theta_1}\right)}, \theta_5 * max(S,0)+\theta_4, \infty\right)$	14
	Saturation excess flow from a store with different degrees of saturation (min-max linear variant)	saturation_8	$flux_{out} = \left[\theta_1 + (\theta_2-\theta_1)\frac{S}{S_{max}}\right] * flux_{in}$	$flux_{out} \leq flux_{in}$	$flux_{out} = \left[\theta_1 + (\theta_2-\theta_1)\frac{S}{S_{max}}\right] * flux_{in}$	45
	Deficit store: saturation excess from a store that has reached maximum capacity	saturation_9	$flux_{out} = \begin{cases} flux_{in}, & if\ S = 0 \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in} * logisticSmoother\_S(S, 0)$	17, 25, 43, 46
	Saturation excess flow from a store with different degrees of saturation (min-max exponential variant)	saturation_10	$flux_{out} = min(\theta_1, \theta_2 + \theta_2 e^{\theta_3 S}) * flux_{in}$		$flux_{out} = min(\theta_1, \theta_2 + \theta_2 e^{\theta_3 S}) * flux_{in}$	39

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Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Saturation excess flow from a store with different degrees of saturation (min exponential variant)	saturation_11	$flux_{out} = \begin{cases} \left(\theta_1 \left[\frac{S-S_{min}}{S_{max}-S_{min}}\right]^{\theta_2}\right) flux_{in}, & if\ S > S_{min} \\ 0, & otherwise \end{cases}$	$flux_{out} \leq flux_{in}$	$flux_{out} = min\left(1, \theta_1 \left[\frac{S-S_{min}}{S_{max}-S_{min}}\right]^{\theta_2}\right) flux_{in} * [1 - logisticSmoother\_S(S, S_{min})]$	23
	Saturation excess flow from a store with different degrees of saturation (min-max linear variant)	saturation_12	$flux_{out} = \frac{\theta_1 - \theta_2}{1 - \theta_2} flux_{in}$	$flux_{out} \geq 0$	$flux_{out} = max\left(0, \frac{\theta_1 - \theta_2}{1 - \theta_2}\right) flux_{in}$	23
	Saturation excess flow from a store with different degrees of saturation (normal distribution variant)	saturation_13	$flux_{out} = flux_{in} * \int_{-\infty}^{\xi} \frac{1}{\sqrt{2\pi}} exp\left[-\frac{\xi^2}{2}\right] d\xi,\ with\ \xi = \frac{log(S/\theta_1)}{log(\theta_1/\theta_2)}$		$flux_{out} = flux_{in} * normcdf\left(\frac{log(\max(0,S)/\theta_1)}{log(\theta_1/\theta_2)}\right)$	42
	Saturation excess flow from a store with different degrees of saturation (two-part exponential variant)	saturation_14	$flux_{out} = flux_{in} \begin{cases} (0.5-\theta_1)^{1-\theta_2} \left(\frac{S}{S_{max}}\right)^{\theta_3}, & if\ \frac{S}{S_{max}} \leq 0.5 - \theta_1 \\ 1 - (0.5-\theta_1)^{1-\theta_2} \left(1 - \frac{S}{S_{max}}\right)^{\theta_3}, & otherwise \end{cases}$		$flux_{out} = \left(\left((0.5-\theta_1)^{1-\theta_2} max\left(0, \frac{S}{S_{max}}\right)^{\theta_3}\right) * \left(\frac{S}{S_{max}} \leq 0.5 - \theta_1\right) + \left(1 - (0.5+\theta_1)^{1-\theta_2} max\left(0, 1 - \frac{S}{S_{max}}\right)^{\theta_3}\right) * \frac{S}{S_{max}} > 0.5 - \theta_1 \right) * flux_{in}$	28
Snow	Snowfall based on temperature threshold	snowfall_1	$flux_{out} = \begin{cases} flux_{in}, & if\ T \leq T_{threshold} \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in} * [logisticSmoother\_T(T, T_{threshold}]$	6, 12, 30, 31, 32, 34, 35, 41, 43, 44, 45
	Snowfall based on a temperature threshold interval	snowfall_2	$flux_{out} = \begin{cases} flux_{in}, & if\ T \leq \theta_1 - \frac{1}{2}\theta_2 \\ flux_{in} * \frac{\theta_1 + \frac{1}{2}\theta_2 - T}{\theta_2}, & if\ \theta_1 - \frac{1}{2}\theta_2 < T < \theta_1 + \frac{1}{2}\theta_2 \\ 0, & if\ T \geq \theta_1 + \frac{1}{2}\theta_2 \end{cases}$		$flux_{out} = min\left(flux_{in}, max\left(0, flux_{in} * \frac{\theta_1 + \frac{1}{2}\theta_2 - T}{\theta_2}\right)\right)$	37
	Rainfall based on temperature threshold	rainfall_1	$flux_{out} = \begin{cases} flux_{in}, & if\ T > T_{threshold} \\ 0, & otherwise \end{cases}$		$flux_{out} = flux_{in} * [1 - logisticSmoother\_T(T, T_{threshold}]$	6, 12, 30, 31, 32, 34, 35, 41, 43, 44, 45

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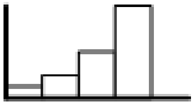
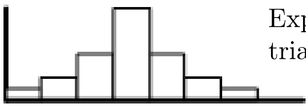
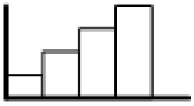
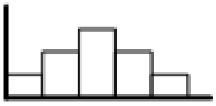
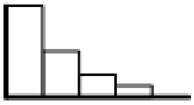
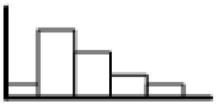
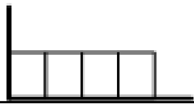
Process	Details	Function name	Constitutive function	Constraints	MARRMoT Code	Model
	Snowfall based on a temperature threshold interval	rainfall_2	$flux_{out} = \begin{cases} 0, \; if\; T \le \theta_1 - \frac{1}{2}\theta_2 \\ flux_{in} * \frac{\theta_1 + \frac{1}{2}\theta_2 - T}{\theta_2}, \; if\; \theta_1 - \frac{1}{2}\theta_2 < T < \theta_1 + \frac{1}{2}\theta_2 \\ flux_{in}, \; if\; T \ge \theta_1 + \frac{1}{2}\theta_2 \end{cases}$		$flux_{out} = min\left(flux_{in}, max\left(0, flux_{in} * \frac{T - \left(\theta_1 - \frac{1}{2}\theta_2\right)}{\theta_2}\right)\right)$	37
	Refreezing of stored melted snow	refreeze_1	$flux_{out} = \begin{cases} \theta_1 * \theta_2 * (T_{threshold} - T), \; if\; T \le T_{threshold} \\ 0, \; otherwise \end{cases}$	$flux_{out} \le \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, max\left(0, \theta_1 * \theta_2 * (T_{threshold} - T)\right)\right)$	37, 44
	Snowmelt from degree-day-factor	melt_1	$flux_{out} = \begin{cases} \theta_1 * (T - T_{threshold}), \; if\; T \ge T_{threshold} \\ 0, \; otherwise \end{cases}$	$flux_{out} \le \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, max\left(0, \theta_1 * (T - T_{threshold})\right)\right)$	6, 12, 30, 31, 32, 34, 35, 37, 43, 44, 45
	Snowmelt at a constant rate	melt_2	$flux_{out} = \begin{cases} \theta_1, \; if\; S \ge 0 \\ 0, \; otherwise \end{cases}$	$flux_{out} \le \frac{S}{\Delta t}$	$flux_{out} = min\left(\frac{S}{\Delta t}, \theta_1\right)$	44
	Glacier melt provided no snow is stored on the ice layer	melt_3	$flux_{out} = \begin{cases} \theta_1 * (T - T_{threshold}), \; if\; T \ge T_{threshold}, \; S_2 = 0 \\ 0, \; otherwise \end{cases}$	$flux_{out} \le \frac{S_1}{\Delta t}$	$flux_{out} = min\left(\frac{S_1}{\Delta t}, max\left(0, \theta_1 * \theta_2 * (T_{threshold} - T)\right)\right) * logisticSmoother\_S(S_2, 0)$	43
Soil moisture	Water rebalance to equal relative storage (2 stores)	soilmoisture_1	$flux_{out} = \begin{cases} \frac{S_2 S_{1,max} - S_1 S_{2,max}}{S_{1,max} + S_{2,max}}, \; if\; \frac{S_1}{S_{1,max}} < \frac{S_2}{S_{2,max}} \\ 0, \; otherwise \end{cases}$		$flux_{out} = \left(\frac{S_2 S_{1,max} - S_1 S_{2,max}}{S_{1,max} + S_{2,max}}\right) * logisticSmoother\_S\left(\frac{S_1}{S_{1,max}}, \frac{S_2}{S_{2,max}}\right)$	33
	Water rebalance to equal relative storage (3 stores)	soilmoisture_2	$flux_{out} = \begin{cases} S_2 \frac{S_1 \left(S_{2,max} + S_{3,max}\right) + S_{1,max}(S_2 + S_3)}{\left(S_{2,max} + S_{3,max}\right)\left(S_{1,max} + S_{2,max} + S_{3,max}\right)}, \; if\; \frac{S_1}{S_{1,max}} < \frac{S_2 + S_3}{S_{2,max} + S_{3,max}} \\ 0, \; otherwise \end{cases}$		$flux_{out} = \left(S_2 \frac{S_1 \left(S_{2,max} + S_{3,max}\right) + S_{1,max}(S_2 + S_3)}{\left(S_{2,max} + S_{3,max}\right)\left(S_{1,max} + S_{2,max} + S_{3,max}\right)}\right) * logisticSmoother\_S\left(\frac{S_1}{S_{1,max}}, \frac{S_2 + S_3}{S_{2,max} + S_{3,max}}\right)$	33

S4 Unit Hydrographs

This section provides details on the implementation of various Unit Hydrographs. An overview of the 7 UHs is given in Table S2. Computational implementation of each UH is given in sections S4.1 to S4.7. Unit Hydrograph files can be found in "./MARRMoT/Models/Unit Hydrograph files/".

Table S2: Overview of Unit Hydrograph schemes implemented in MARRMoT

File name	**Inputs**	**Diagram**	**Description**	**In model ...**
uh_1_half	1: amount to be routed 2: time base 3: Δt		Exponentially increasing scheme	7
uh_2_full	1: amount to be routed 2: time base (time is doubled inside the function) 3: Δt		Exponential triangular scheme	7
uh_3_half	1: amount to be routed 2: time base 3: Δt		Triangular scheme: linearly increasing	13, 15, 21, 26 34
uh_4_full	1: amount to be routed 2: time base 3: Δt		Triangular scheme: linearly increasing and decreasing	0 (template), 16, 37, nn (example)
uh_5_half	1: amount to be routed 2: time base 3: Δt		Exponentially decreasing scheme	5
uh_6_gamma	1: amount to be routed 2: gamma parameter [-] 3: time for flow to reduce by factor e [d] 4: length of time series		Gamma function-based	40
uh_7_uniform	1: amount to be routed 2: time base 3: Δt		Uniform distribution	39

S4.1 Code: uh_1_half

This section provides the computational implementation of a unit hydrograph with an increasing exponential distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_1_half
References	E.g. GR4J Perrin et al. (2003)

```
function [ out,UH ] = uh_1_half( in, d_base, delta_t )
%uh_1_half Unit Hydrograph [days] with half a bell curve. GR4J-based
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
%
%   Inputs
%   in      - volume to be routed
%   d_base  - time base of routing delay [d]
%   delta_t - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period x4.
%   Percentage of input returned only increases.
%   I.e. d_base = 3.8 [days], delta_t = 1:
%   UH(1) = 0.04  [% of inflow]
%   UH(2) = 0.17
%   UH(3) = 0.35
%   UH(4) = 0.45

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
if delay == 0; delay = 1; end        % any value below t = 1 means no delay,
                                     % but zero leads to problems
tt = 1:ceil(delay);                  % Time series for which we need UH
                                     % ordinates [days]
```

```

%%EMPTIES
SH = zeros(1,length(tt)+1); SH(1) = 0;
UH = zeros(1,length(tt));

%%UNIT HYDROGRAPH
for t = tt
    if     t <  delay; SH(t+1) = (t./delay).^(5./2);
    elseif t >= delay; SH(t+1) = 1;
    end

    UH(t) = SH(t+1)-SH(t);
end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.2 Code: uh_2_full

This section provides the computational implementation of a unit hydrograph with an exponential triangular distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_2_full
References	E.g. GR4J Perrin et al. (2003)

```
function [ out, UH ] = uh_2_full( in,d_base,delta_t )
%uh_2_full Unit Hydrograph [days] with a full bell curve.
    GR4J-based
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and
    distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details
    .
%
%   Inputs
%   in      - volume to be routed
%   d_base  - time base of routing delay [d]
%   delta_t - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time
    period 2*x4.
```

```
%   Percentage of input returned goes up (till x4), then down again.
%   I.e. d_base = 3.8 [days], delta_t = 1:
%   UH(1) = 0.02  [% of inflow]
%   UH(2) = 0.08
%   UH(3) = 0.18
%   UH(4) = 0.29
%   UH(5) = 0.24
%   UH(6) = 0.14
%   UH(7) = 0.05
%   UH(8) = 0.00

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
tt = 1:2*ceil(delay); % time series for which we need UH ordinates [days]

%%EMPTIES
SH = zeros(1,length(tt)+1); SH(1) = 0;
UH = zeros(1,length(tt));

%%UNIT HYDROGRAPH
for t = tt
    if      (t <= delay)
        SH(t+1) = 0.5*(t./delay).^(5./2);
    elseif (t > delay) && (t < 2*delay);
        SH(t+1) = 1-0.5*(2-t./delay).^(5./2);
    elseif (t >= 2*delay);
        SH(t+1) = 1;
    end

    UH(t) = SH(t+1)-SH(t);
end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.3 Code: uh_3_half

This section provides the computational implementation of a unit hydrograph with an linearly increasing distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_3_half
References	E.g. FLEX-Topo Savenije (2010)

```
function [ out,UH ] = uh_3_half( in, d_base, delta_t )
%uh_3_half Unit Hydrograph [days] with half a triangle (linear)
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
%
%   Inputs
%   in      - volume to be routed
%   d_base  - time base of routing delay [d]
%   delta_t - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period delay.
%   Percentage of input returned only increases.
%   I.e. d_base = 3.8 [days], delta_t = 1:
%   UH(1) = 0.04  [% of inflow]
%   UH(2) = 0.17
%   UH(3) = 0.35
%   UH(4) = 0.45

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
if delay == 0; delay = 1; end   % any value below t = 1 means no delay,
                                % but zero leads to problems
tt = 1:ceil(delay);             % time series for which we need UH
                                % ordinates [days]
```

```

%%UNIT HYDROGRAPH
% The area under the unit hydrograph by definition sums to 1. Thus the area
% is S(t=0 to t = delay) t*[ff: fraction of flow to move per time step] dt
% Analytical solution is [1/2 * t^2 + c]*ff, with c = 0. Thus the fraction
% of flow step size is:
ff = 1/(0.5*delay^2);

%%EMPTIES
UH = zeros(1,length(tt));

%%UNIT HYDROGRAPH
for t = 1:length(tt)
    if t <= delay
        UH(t) = ff.*(0.5*t^2 - 0.5*(t-1)^2);
    else
        UH(t) = ff.*(0.5*delay^2 - 0.5*(t-1)^2);
    end
end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.4 Code: uh_4_full

This section provides the computational implementation of a unit hydrograph with an linear triangular distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_4_full
References	E.g. HBV-96
citeLindstrom1997	

```
function [ out,UH ] = uh_4_full( in, d_base, delta_t )
%uh_4_half Unit Hydrograph [days] with a triangle (linear)
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
```

```
%
%   Inputs
%   in      - volume to be routed
%   d_base  - time base of routing delay [d]
%   delta_t - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period delay.
%   Percentage runoff goes up, peaks, and goes down again.
%   I.e. d_base = 3.8 [days], delta_t = 1:
%   UH(1) = 0.14  [% of inflow]
%   UH(2) = 0.41
%   UH(3) = 0.36
%   UH(4) = 0.09

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
if delay == 0; delay = 1; end        % any value below t = 1 means no delay,
                                     % but zero leads to problems
tt = 1:ceil(delay);                  % time series for which we need UH
                                     % ordinates [days]

%%UNIT HYDROGRAPH
% The area under the unit hydrograph by definition sums to 1. Thus the area
% is S(t=0 to t = delay) t*[ff: fraction of flow to move per time step] dt
% Analytical solution is [1/2 * t^2 + c]*ff, with c = 0.
% Here, we use two half triangles t make one big one, so the area of half a
% triangle is 0.5. Thus the fraction of flow step size is:
ff  = 0.5/(0.5*(0.5*delay)^2);
d50 = 0.5*delay;

%%TRIANGLE FUNCTION
tri = @(t) max(ff.*(t-d50).*sign(d50-t)+ff.*d50,0);

%%EMPTIES
```

```
UH = zeros(1,length(tt));

%%UNIT HYDROGRAPH
for t = 1:length(tt)
    UH(t) = integral(tri,t-1,t);
end

%%ENSURE UH SUMS TO 1
tmp_diff    = 1-sum(UH);
tmp_weight = UH./sum(UH);
UH          = UH + tmp_weight.*tmp_diff;

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.5 Code: uh_5_half

This section provides the computational implementation of a unit hydrograph with an decreasing exponential distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_5_half
References	E.g. IHACRES Littlewood et al. (1997); Croke and Jakeman (2004)

```
function [ out,UH ] = uh_1_half( in, d_base, delta_t )
%uh_1_half Unit Hydrograph [days] with half a bell curve. GR4J-based
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
%
%   Inputs
%   in       - volume to be routed
%   d_base   - time base of routing delay [d]
%   delta_t  - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period x4.
%   Percentage of input returned only increases.
%   I.e. d_base = 3.8 [days], delta_t = 1:
```

```
%    UH(1) = 0.04   [% of inflow]
%    UH(2) = 0.17
%    UH(3) = 0.35
%    UH(4) = 0.45

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
if delay == 0; delay = 1; end        % any value below t = 1 means no delay,
                                     % but zero leads to problems
tt = 1:ceil(delay);                  % Time series for which we need UH
                                     % ordinates [days]

%%EMPTIES
SH = zeros(1,length(tt)+1); SH(1) = 0;
UH = zeros(1,length(tt));

%%UNIT HYDROGRAPH
for t = tt
    if      t <  delay; SH(t+1) = (t./delay).^(5./2);
    elseif t >= delay; SH(t+1) = 1;
    end

    UH(t) = SH(t+1)-SH(t);
end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.6 Code: uh_6_gamma

This section provides the computational implementation of a unit hydrograph with a gamma distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_6_gamma
References	E.g. SMAR O'Connell et al. (1970); Tan and O'Connor (1996)

```matlab
function [ out,UH,frac_routing_beyond_time_series ] = ...
                                                uh_6_gamma( in,n,k,t_end,delta_t )
%uh_6_gamma Unit Hydrograph [days] from gamma function.
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
%
%   Inputs
%   n        = shape parameter [-]
%   k        = time delay for flow reduction by a factor e [d]
%   t_end    = length of time series [d]
%   delta_t = time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period delay.
%   Percentage of input returned only decreases.
%   I.e. n = 1, k = 3.8 [days], delta_t = 1:
%   UH(1) = 0.928  [% of inflow]
%   UH(2) = 0.067
%   UH(3) = 0.005
%   UH(4) = 0.000

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single value.'); end

%%TIME STEP SIZE
tmax = t_end/delta_t;
tt    = 1:tmax;            % time series for which we need UH ordinates [days]

%%EMPTIES
UH_full   = zeros(1,length(tt));
frac_routing_beyond_time_series = 0;

%%UNIT HYDROGRAPH
% The Unit Hydrograph follows a gamma distribution. For a given
% delay time, the fraction of flow per time step is thus
```

```
the integral of
% t-1 to t of the gamma distrubtion. The curve has range [0,Inf>.
% We need to choose a point at which to cap the integration, but this
% depends on the parameters n & k, and the total time step. We choose the
% cutoff point at the time step where less than 0.1% of the peak flow
% is still on route.

%%Unit hydrograph
for t = 1:length(tt)
    UH_full(t) = integral(@(x) 1./(k.*gamma(n)).*(x./k).^(n-1).* ...
                                        exp(-1.*x./k),(t-1)*delta_t,t*delta_t);
end

%%Find cutoff point where less than 0.1% of the peak flow is being routed
[max_val,max_here] = max(UH_full);
end_here = find(UH_full(max_here:end)./max_val<0.001,1) + max_here;

%%Take action depending on whether the distribution function exceeds the
%%time limit or not
if ~isempty(end_here)
    %%Construct the Unit Hydrograph
    UH = UH_full(1:end_here);

    %%Account for the truncated part of the full UH.
    % find probability mass to the right of the cut-off point
    tmp_excess = 1-sum(UH);

    % find relative size of each time step
    tmp_weight = UH_full(1:end_here)./sum(UH_full(1:end_here));

    % distribute truncated probability mass proportionally to all elements
    % of the routing vector
    UH = UH+tmp_weight.*tmp_excess;
```

```

else
    %%Construct the Unit Hydrograph
    UH = UH_full;

    %%The UH is longer than the provided time series length. Track the
    %%percentage of flow that is routed beyond the simulation duration
    frac_routing_beyond_time_series = 1-sum(UH);

end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S4.7 Code: uh_7_uniform

This section provides the computational implementation of a unit hydrograph with a uniform distribution of flows.

File location	./MARRMoT/Models/Unit Hydrograph files/uh_7_uniform
References	E.g. MCRM ?Moore and Bell (2001)

```
function [ out,UH ] = uh_7_uniform( in, d_base, delta_t )
%uh_7_uniform Unit Hydrograph [days] with uniform spread
%
% Copyright (C) 2018 W. Knoben
% This program is free software (GNU GPL v3) and distributed WITHOUT ANY
% WARRANTY. See <https://www.gnu.org/licenses/> for details.
%
%   Inputs
%   in      - volume to be routed
%   d_base  - time base of routing delay [d]
%   delta_t - time step size [d]
%
%   Unit hydrograph spreads the input volume over a time period delay.
%   I.e. d_base = 3.8 [days], delta_t = 1:
%   UH(1) = 0.26  [% of inflow]
```

```
%    UH(2) = 0.26
%    UH(3) = 0.26
%    UH(4) = 0.22

%%INPUTS
if any(size(in)) > 1; error('UH input should be a single
    value.'); end

%%TIME STEP SIZE
delay = d_base/delta_t;
tt = 1:ceil(delay); % time series for which we need UH
    ordinates [days]

%%EMPTIES
UH = NaN.*zeros(1,length(tt));

%%FRACTION FLOW
ff = 1/delay; % fraction of flow per time step

%%UNIT HYDROGRAPH
for t=1:ceil(delay)
    if t < delay
        UH(t) = ff;
    else
        UH(t) = mod(delay,t-1)*ff;
    end
end

%%DISPERSE VOLUME
out = in.*UH;

end
```

S5 Parameter ranges

Each model function in MARRMoT is accompanied by a file that specifies suitable sampling ranges for each parameter used in the model, that could be applied if the user chooses to pair MARRMoT with a calibration or parameter sampling procedure. This section gives the reasoning behind our choices of parameter ranges used within MARRMoT.

S5.1 Model-specific ranges versus generalised process-specific ranges

There are two different approaches to determining parameter ranges for model calibration or parameter sampling studies: (1) make a choice for appropriate parameter ranges per model, based on previous applications of the model, or (2) try to make consistent choices for all models based on literature (e.g. ensure that all 'slow' linear reservoirs, regardless of which model they are part of, have the same limits for the drainage time scale parameter). Generalization of parameter ranges across models is difficult because models use different flux formulations and thus different parameter values might be appropriate, even if the fluxes are intended to represent the same hydrologic process. On the other hand, using model-specific parameter ranges based on earlier studies might limit a model's potential. Especially if the model has only been applied to a small number of places, published 'appropriate' parameter ranges might also reflect the climate or catchment characteristics of the few study catchments the model has been applied to. MARRMoT is intended as a model comparison framework. We thus attempt to generalize parameter ranges across all models in the framework, to facilitate fair comparison of different models. We try to err on the side of caution and intentionally set these ranges wide. Table S3 shows the parameter ranges used in MARRMoT and specifies in which model(s) each parameter range is used.

Table S3: Parameter ranges used in MARRMoT

Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Snow							
Threshold temperature for snowfall (and melt, if not specified otherwise) $[^oC]$	Table S4	Table S4	-3	5	Kienzle (2008); Kollat et al. (2012)		6, 12, 30, 31, 32, 34, 35, 37, 43, 44, 45
Threshold interval width for snowfall $[^oC]$	0	7	0	17	Kienzle (2008)	0 is a physical limit	37
Threshold temperature for melt $[^oC]$			-3	3		Not easy to find any interval. Temperature for melt tends be treated as constant at 0	37, 43, 44
Degree-day-factor for snow or ice melt $[mm/^oC/d]$	0	Table S5	0	20		0 is a physical limit	6, 12, 30, 31, 32, 34, 35, 37, 41, 43, 44, 45
Water holding content of snow pack [-]	0	0.8	0	1	Kollat et al. (2012)	[0,1] are physical limits	37, 44
Refreezing factor of retained liquid water [-]	0	1	0	1		[0,1] are physical limits	37, 44 (given as fraction [0,1] of degree-day-factor)
Maximum melt rate due to ground-heat flux [mm/d]	0	2	0	2	Schaefli et al. (2014)		44

continued . . .

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Interception							
Maximum store depth [mm]	0	Table S6	0	5	Chiew and McMahon (1994); Gerrits (2010)	0 is a physical limit. Gerrits (2010) (table 1.1) reports 3.8mm as maximum value used out of 15 studies. Chiew and McMahon (1994) (table 3) report 5.6mm as a maximum value for 28 catchments	2, 13, 15, 16, 18, 22, 23, 26, 34, 36, 39, 42, 44, 45
Maximum intercepted fraction of precipitation [-]	0	0.42	0	1	Gerrits (2010)	[0,1] are physical limits. Gerrits (2010) (table 1.1) reports 42% as maximum intercepted fraction out of 15 studies	8, 23, 32, 35, 45
Seasonal variation in LAI as fraction of mean [-]			0	1		0 is a physical limit	22
Timing of maximum Leaf Area Index [d]			1	365		Refers to days in a normal calendar year	22, 32, 35
Surface depression							
Maximum surface area contributing to store [-]	0	1	0	1		[0,1] are physical limits	36, 45
Maximum store depth [mm]	0	Table S7	0	50	Chiew and McMahon (1994)	0 is physical limit. 50 is recommended in Chiew and McMahon (1994)	36, 45

continued . . .

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Filling parameter [-]	1	1	0.99	1	Chiew (1990); Porter and McMahon (1971)	Controls the shape of the depression store inflow flux but is usually set at 1 because no studies are (were?) available about how a depression store fills	36
Infiltration							
Maximum loss [mm]	0	400	0	600	Chiew et al. (2002)	Fig 11.11a shows calibrated parameter values for 339 catchments. Pattern indicates that limit was set at 400	18, 36
Loss exponent [-]	0	12	0	15	Chiew et al. (2002)	Fig 11.11a shows calibrated parameter values for 339 catchments. Pattern indicates that limit was set at 10	18, 36
Maximum infiltration rate [mm/d]	Table S8	Table S8	0	200		Infiltration rates can be very high. However, to have a practical effect on modelling, (i.e. generate infiltration excess flow), Inf_rate < P(t). In the context of a follow-up study, Inf_rate is capped at 200mm/d because the maximum daily P in the study area is 200mm/d.	15, 20, 23, 40, 44

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Infiltration decline non-linearity parameter [-]			0	5	Sivapalan et al. (1996)	Very difficult to find information for (original paper mentions nothing)	23, 43
Evaporation							
Plant-controlled maximum rate [mm/d]	5	24.5	0	20	Chiew and McMahon (1994)	Although the study reports an upper value of 24.5, the recommended range is capped at 20 (paper appendix)	20, 36
Wilting point as fraction of Soil moisture capacity [-]	0.1	0.25	0.05	0.95	Son and Sivapalan (2007)	0 is a physical limit but can break model equations through "divide-by-zero" errors. 1 is a physical limit	3, 4, 8, 9, 10, 12, 14, 15, 16, 19, 20, 21, 26, 31, 32, 34, 35, 37, 44
Moisture constrained rate parameter [-]			0	1		[0,1] are physical limits	15
Forest fraction for separate soil/vegetation evap [-]	0	1	0.05	0.95		[0,1] are physical limits, but using these limits can result in divide-by-zero-errors in certain fluxes	3, 4, 8, 16
Phenology: minimum temperature where transpiration stops [^oC]	-5	-5	0	-10	Ye et al. (2012)		35

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Phenology: maximum temperature above which transpiration fully utilizes Ep [^oC]	10	10	1	20	Ye et al. (2012)	The setup of minimum and maximum temperature used in Ye et al. (2012) is here changed to a minimum temperature + temperature range (Tmax = Tmin + Trange) to avoid overlap in parameter values	35
Evaporation reduction with depth coefficient [-]	0.083	1	0	1	Penman (1950); Tan and O'Connor (1996)	[0,1] are physical limits	17, 23, 25, 40
Shape parameter for evaporation reduction in a deficit store [-]			0	1	Moore and Bell (2001)	This uses a sigmoid function to determine a fraction of Ep to evaporate. Values >1 make the transition very steep	39
Evaporation non-linearity coefficient [-]			0	10	Sivapalan et al. (1996)	Very difficult to find information for. Assumption made to be in line with other non-linearity coefficients.	23, 43
Soil moisture							
Maximum store depth [mm]	1	Table S9	1	2000		0 is a physical limit	1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 26, 27, 28, 29, 30, 31, 32, 33, 34, 35, 36, 37, 38, 39, 40, 41, 42, 43, 44, 45, 46

continued . . .

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Capillary rise [mm/d]	0	Table S10	0	4		0 is a physical limit SMHI gives a default value of 1 mm/d for use with HBV. We use a wider range here	13, 15, 37, 38
Percolation rate [mm/d]	0	Table S11	0	20	Bethune et al. (2008)	Some modelling studies report very large percolation rates (100 mm/d). Bethune et al. (2008) report 11mm/d from field observations.	21, 26, 34, 37, 39, 44, 45
Percolation fraction [-]	0.013	0.533	0	1	Ye et al. (2012) (Table 1)	[0,1] are physical limits	14, 22, 23, 24, 27, 30, 31, 32, 35, 45
Recharge nonlinearity [-]	0	7	0	10	Kollat et al. (2012)	Also seen as a soil depth distribution	5, 22, 33, 37
Soil depth distribution [-]	0	Table S12	0	10		For cases where the soil depth is not considered constant. Most studies limit this to 0-2.5 but this seems based on a single source (Wagener et al., 2004) which is UK only. Thus we use a wider range here	2, 13, 15, 21, 22, 26, 28, 29, 34
Porosity [-]	0.35	0.5	0.05	0.95	Son and Sivapalan (2007)	[0,1] are theoretical physical limits, but no (0) porosity and full (1) porosity are not sensible: there would be no soil moisture or soil respectively	10, 19

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Gamma distribution for topographic indices - phi [-]	0.4	3.5	0.1	5	Clark et al. (2008)		14
Gamma distribution for topographic indices - chi [-]	2	5	1	7.5	Clark et al. (2008)		14
Fraction area with permeable soils [-]			0	1	Crooks and Naden (2007)	[0,1] are physical limits	46
Fraction area with semi-permeable soils [-]			0	1	Crooks and Naden (2007)	[0,1] are physical limits	46
Fraction area with impermeable soils [-]			0	1	Crooks and Naden (2007)	[0,1] are physical limits	46
Variable contributing area scaling [-]			0	5	Sivapalan et al. (1996)	Very difficult to find information about this. Assumption made	23
Variable contributing area non-linearity [-]					Sivapalan et al. (1996)	See: **Soil depth distribution** above	23
Fraction of D50 that is D16 [-]			0.01	0.99		Note: re-writing of D16 parameter in Fukushima (1988)	42
Variable contributing area equation inflection point [-]	-0.5	0.5	-0.5	0.5	Jayawardena and Zhou (2000)		28
Groundwater							
Leakage coefficient [-]	0.07	0.13	0	0.5	Chiew and McMahon (1994)	0 is physical limit. 0.5 is recommended in the paper's appendix	36
Leakage rate [mm/d]						See: **Percolation rate** above	

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Level compared to channel level [mm]	-2.8	3.9	-10	10	Chiew and McMahon (1994)	Range recommended in appendix of the paper	36
Base flow rate at no deficit [mm/d]	0	201.6	0.1	200	Beven (1997)	Based on Table 2 (Beven, 1997)	14, 23
Baseflow deficit scaling parameter [-]			0	1		[0,1] are physical limits	14, 23
Flow distribution							
Interflow and saturation excess [-]	0	1	0	1		[0,1] are physical limits	18, 36
Preferential recharge [-]	0	2	0	1	Chiew and McMahon (1994)	0 is a physical limit. Later paper sets max limit to 1	18, 25, 36, 46
Surface/groundwater division [-]			0	1		[0,1] are physical limits	13, 17, 33
Fast and slow flow [-]	0	1	0	1		[0,1] are physical limits	21, 26, 29, 34, 46
Groundwater recharge and interflow [-]	0.05	0.3	0	1	Son and Sivapalan (2007)	[0,1] are physical limits	10, 11, 20, 40
Infiltration and direct runoff [-]	0.161	0.422	0	1	Tan and O'Connor (1996)	[0,1] are physical limits	40
Impervious and infiltration area [-]			0	1		[0,1] are physical limits	28, 33, 45
Contributing area to overland flow [-]			0	1		[0,1] are physical limits	39, 45
Tension water and free water [-]			0	1		[0,1] are physical limits	33
Threshold for overland flow generation [-]	0	<1	0	0.99	Nielsen and Hansen (1973)	[0,1] are physical limits	41

continued . . .

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Threshold for overland flow generation [-]	0	<1	0	0.99	Nielsen and Hansen (1973)	[0,1] are physical limits	41
Channel and land division [-]			0	1		[0,1] are physical limits	42
Throughfall/stem flow division [-]			0	1		[0,1] are physical limits	42
Glacier/non-glacier precipitation [-]			0	1		[0,1] are physical limits	43
Flow time scale and shape							
Fast reservoir time scale [d-1]	0.05	Table S13	0	1		0 is a physical limit	12, 21, 24, 26, 28, 29, 30, 31, 32, 33, 34, 35, 37, 39, 41, 42, 43, 44, 46
Slow reservoir time scale [d-1]	0.01	Table S14	0	1		0 is a physical limit	2, 3, 4, 6, 8, 10, 13, 15, 16, 17, 18, 19, 20, 21, 22, 24, 25, 26, 28, 29, 30, 31, 32, 33, 34, 35, 37, 39, 40, 41, 42, 43, 44, 46
Flow non-linearity Sx̂ [-]	0	Table S15	1	5			4, 9, 10, 11, 16, 19, 22, 23, 37, 39, 42, 44, 45
Flow reduction (S/X) [mm]	5	40	1	50	Son and Sivapalan (2007)		9
Exponential shape parameter [mm-1]			0	2	Moore and Bell (2001)	Very difficult to find documentation for	39

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Routing							
Routing delay to fast flow [d]	0	1	1	5	Fenicia et al. (2008)		5, 21, 26, 34
Routing delay to slow flow [d]	0	8	1	15			5, 7, 21, 26, 34
Routing delay [d]	1	Table S16	1	120	Kollat et al. (2012)	1 is the limit (water shouldnâĂŹt speed up). 120 because it seems very high	13, 15, 16, 21, 37, 39, 40
Routing store depth [mm]	1	300	1	300	Perrin et al. (2003)		7, 20, 39, 45
Gamma function, number of Nash cascade reservoirs [-]	0.75	9.79	1	10	**?**	0 would mean no routing, so slightly above that	40
Water exchange parameters							
Coefficient 1 [-]	0.005	0.54	0	1	Chiew and McMahon (1994)	Although the study only reports values up to 0.54, an upper range of 1 is recommended in the studyâĂŹs appendix	36
Coefficient 2 [-]	0.01	0.29	0	1	Chiew and McMahon (1994)	Although the study only reports values up to 0.29, an upper range of 1 is recommended in the studyâĂŹs appendix	36
Coefficient 3 [-]	0	13	0	100	Chiew and McMahon (1994)	Although the study only reports values up to 13, an upper range of 100 is recommended in the studyâĂŹs appendix	36

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Description	Min(lit)	Max(lit)	Min(used)	Max(used)	Reference(s)	Notes	Model
Water exchange coefficient [mm/d]	-10	14	-10	15	Perrin et al. (2003); Santos et al. (2017)	Parameter x2 in GR4J model	7

Table S4: Literature-based ranges for snowmelt parameter ”threshold temperature for snowfall”

Threshold temperature for snowfall $[^oC]$	**Min**	**Max**
Table 2 in Seibert (1997)	-2.5	2.5
Table 1 in Kollat et al. (2012)	-3	3
Table 2 in Kienzle (2008) Note: always coupled with a snow interval [10,17]	1.1	4.5
Table A3 in Seibert and Vis (2012)	-1.5	2.5

Table S5: Literature-based ranges for snowmelt parameter ”degree-day-factor”

Degree-day factor for snowmelt $[mm/C/d]$	**Min**	**Max**
Table 2 in Seibert (1997)	1	10
Table 1 in Kollat et al. (2012)	0	20
Table A3 in Seibert and Vis (2012)	1	10

Table S6: Literature-based ranges for interception parameter ”maximum interception capacity”

Interception bucket [mm]	**Min**	**Max**
Figure 11.11a in Chiew et al. (2002)	0	5
Table 3 in Chiew and McMahon (1994)	0.5	5.6
Table 1.1 in Gerrits (2010)	0	3.8
Table 2 in Son and Sivapalan (2007)		0.4

Table S7: Literature-based ranges for depression parameter ”maximum depression capacity”

Depression bucket [mm]	**Min**	**Max**
Table 3 in Chiew and McMahon (1994)	1	100
Table 1 in Amoah et al. (2013)	5	110

Table S8: Literature-based ranges for infiltration parameter ”maximum infiltration rate”

Infiltration rate	**Min**	**Max**
Figure 2 in Assouline (2013) [mm/d]	40	100
Table 3.3 in Jones (1997) [mm/h]	6	76
Table 3 in Cerdà (1996) [mm/h]	50	770

Table S9: Literature-based ranges for soil moisture parameter ”maximum soil moisture capacity”

Soil moisture bucket [mm]	**Min**	**Max**
Figure 11.11b in Chiew et al. (2002)	0	500
Table 3 in Chiew and McMahon (1994)	65	400
Table 2 in Seibert (1997)	50	500
Table 1 in Rusli et al. (2015)	100	800
Table 1 in Kollat et al. (2012)	0	2000
Table A3 in Seibert and Vis (2012)	50	500
Table 3 in Sun et al. (2015)	1	500

Table S10: Literature-based ranges for capillary rise parameter ”maximum capillary rise rate”

Capillary rise [mm/d]	**Min**	**Max**
Table 1 in Rusli et al. (2015)	0.1	1
Default value in SMHI (2004)	1	1
Figure 3 in Bethune et al. (2008)	0	0.06

Table S11: Literature-based ranges for percolation parameter ”maximum percolation rate”

Percolation rate [mm/d]	**Min**	**Max**
Table 2 in Seibert (1997)	0	6
Table 1 in Rusli et al. (2015)	0.1	5
Table 1 in Kollat et al. (2012)	0	100
Figure 3 in Bethune et al. (2008)	0	10.4
Table A3 in Seibert and Vis (2012)	0	3

Table S12: Literature-based ranges for soil moisture parameter ”soil depth distribution non-linearity”

Soil depth distribution [-]	**Min**	**Max**
Table 3 in Sun et al. (2015)	0	2
Figure 9 in Lamb (1999)	0	2.5
Table 4 in Bulygina et al. (2009)	0	2.5
Figure 4.12 in Wagener et al. (2004)	0	2
Page 700 in Sivapalan and Woods (1995)		4.03
Figure 4 in Huang et al. (2003) Note: estimated values, 97% $<$ 6	0	11.5

Table S13: Literature-based ranges for flow parameter ”fast flow time scale”

Fast flow time scale $[d^{-1}]$	**Min**	**Max**
Table 2 in Seibert (1997)	0.05	0.5
Table 1 in Rusli et al. (2015)	0.05	0.8
Table 1 in Kollat et al. (2012)	0.01	1
Table A3 in Seibert and Vis (2012)	0.01	0.4
Table 3 in Sun et al. (2015)	0.5	1.2

Table S14: Literature-based ranges for flow parameter ”slow flow time scale”

Slow flow time scale $[d^{-1}]$	**Min**	**Max**
Figure 11.11b in Chiew et al. (2002)	0	0.3
Table 2 in Son and Sivapalan (2007)	2.40E-05	0.1
Table 2 in Seibert (1997)	0.001	0.1
Table 1 in Rusli et al. (2015)	0.0005	0.1
Table 1 in Kollat et al. (2012)	0.00005	0.05
Table A3 in Seibert and Vis (2012)	0.001	0.15
Table 3 in Sun et al. (2015)	0.001	0.5

Table S15: Literature-based ranges for flow parameter ”flow non-linearity”

Flow non-linearity	**Min**	**Max**
Table 3 in Lidén and Harlin (2000) âĂŞ non-linearity shape = S^{1+var}	0	3
Table 1 in Son and Sivapalan (2007) âĂŞ non-linearity shape = $S^{1/var}$	0.45	0.5
Table 3 in Jothityangkoon et al. (2001)	0.5	0.5

Table S16: Literature-based ranges for routing parameter ”routing delay”

Routing delay [d]	**Min**	**Max**
Table 2 in Seibert (1997)	1	5
Table 1 in Kollat et al. (2012)	24	120
Table 3 in Lidén and Harlin (2000)	1	4
Table 1 in Perrin et al. (2003)	0.5	4
Table A3 in Seibert and Vis (2012)	1	7
Table 2 in Atkinson et al. (2003) Note: converted from a flow speed of 0.5m/s and catchment area of $47km^2$		<1
Table 3 in Goswami and O'Connor (2010)	12	36
Table 2 in Vinogradov et al. (2011) Note: approximated from flow velocities and catchment sizes	0.01	4

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