Climate and the local water balance

Healthy soil, South Dakota

Today's goals

- Understanding moisture-dependency of key fluxes
- Construction of a simple hydrological model
- Interplay between climate and water balance

Evapotranspiration and catchment hydrology

In many catchments, more water leaves the catchment via evaporation than via streamflow.

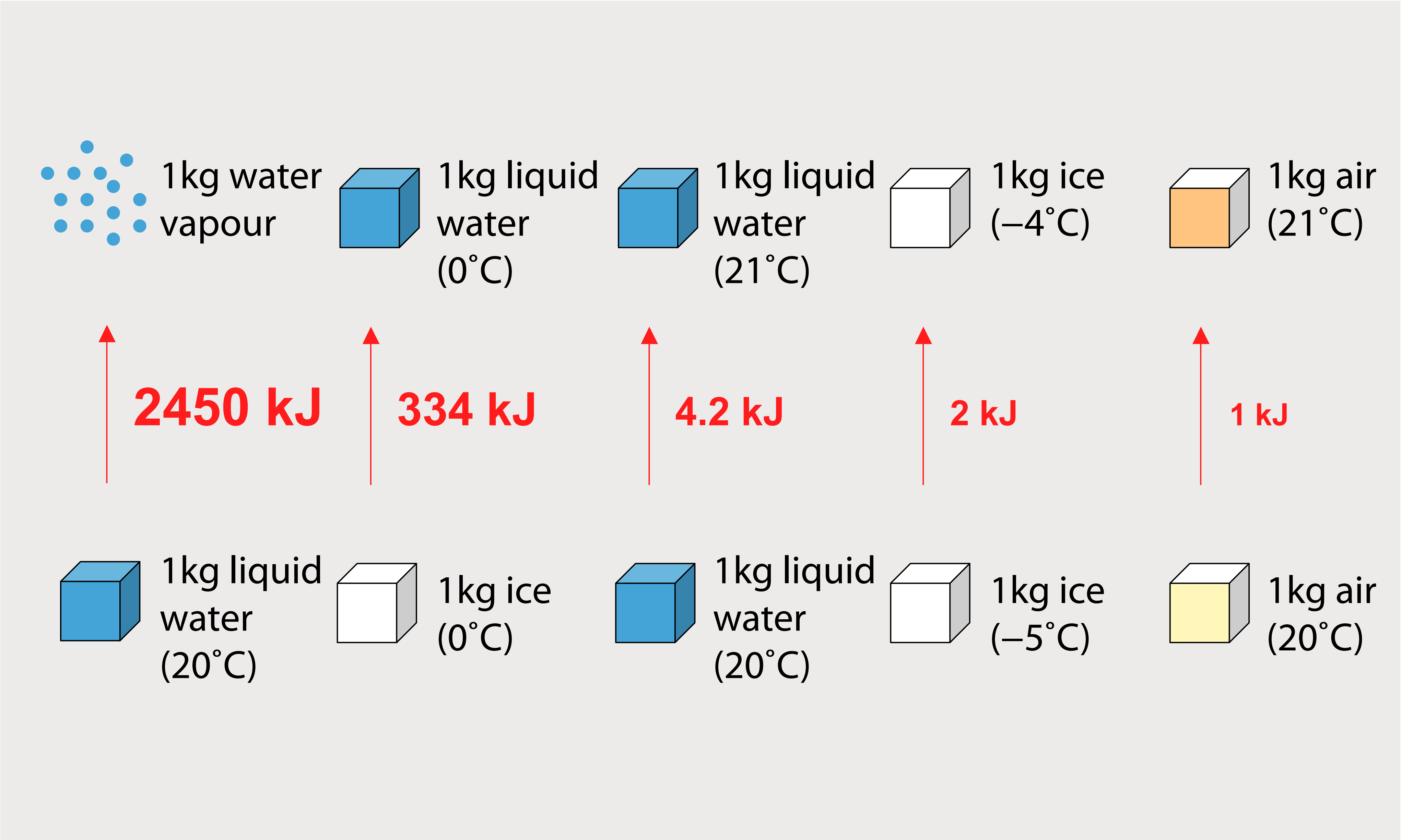
Streamflow is often storage-driven, leading to faster recession rates in summer compared to winter.

Many long-term changes in runoff characeristics can be attributed to changes in evaporation.

Integral part of hydrological models.

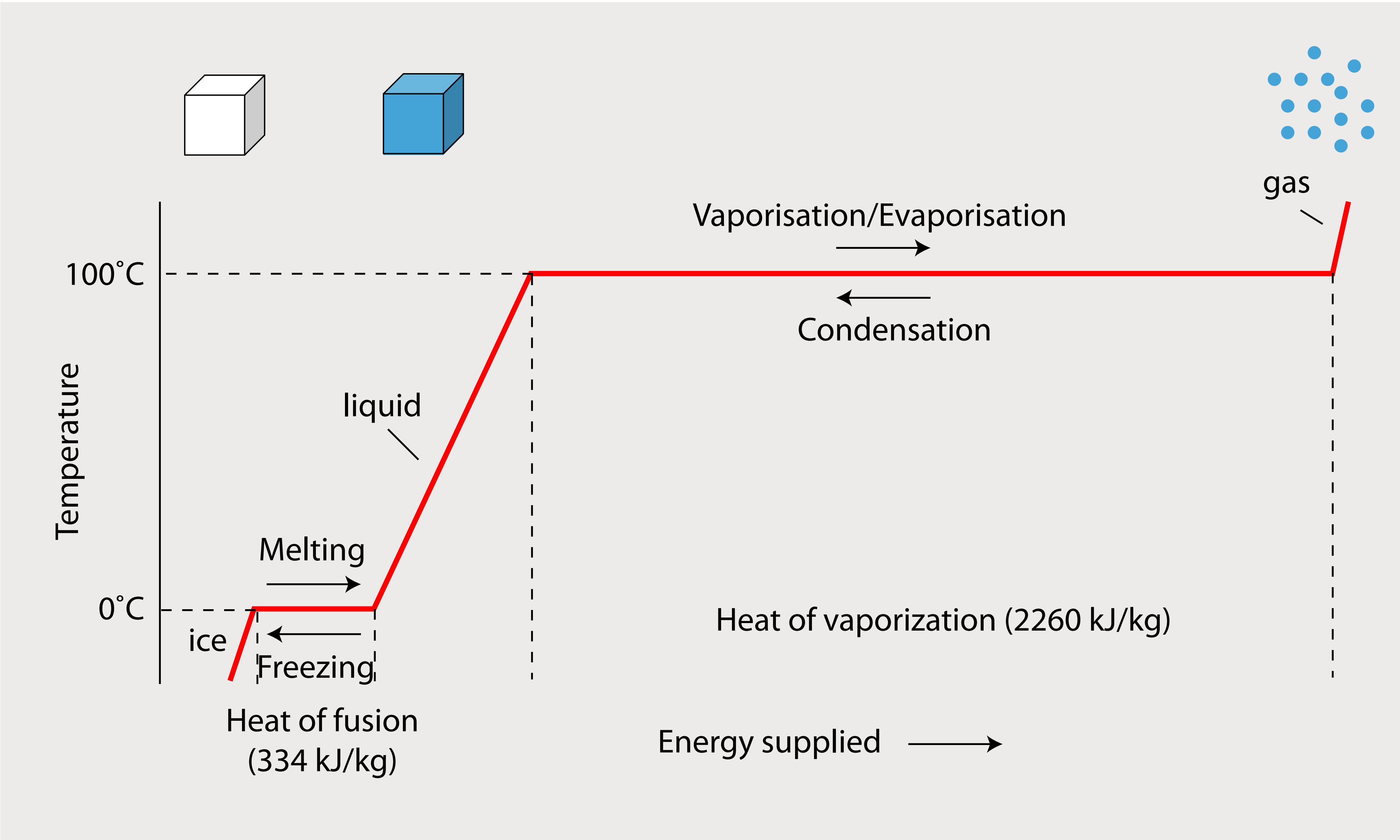
Rietholzbach, Switzerland

Energy associated with water phase changes

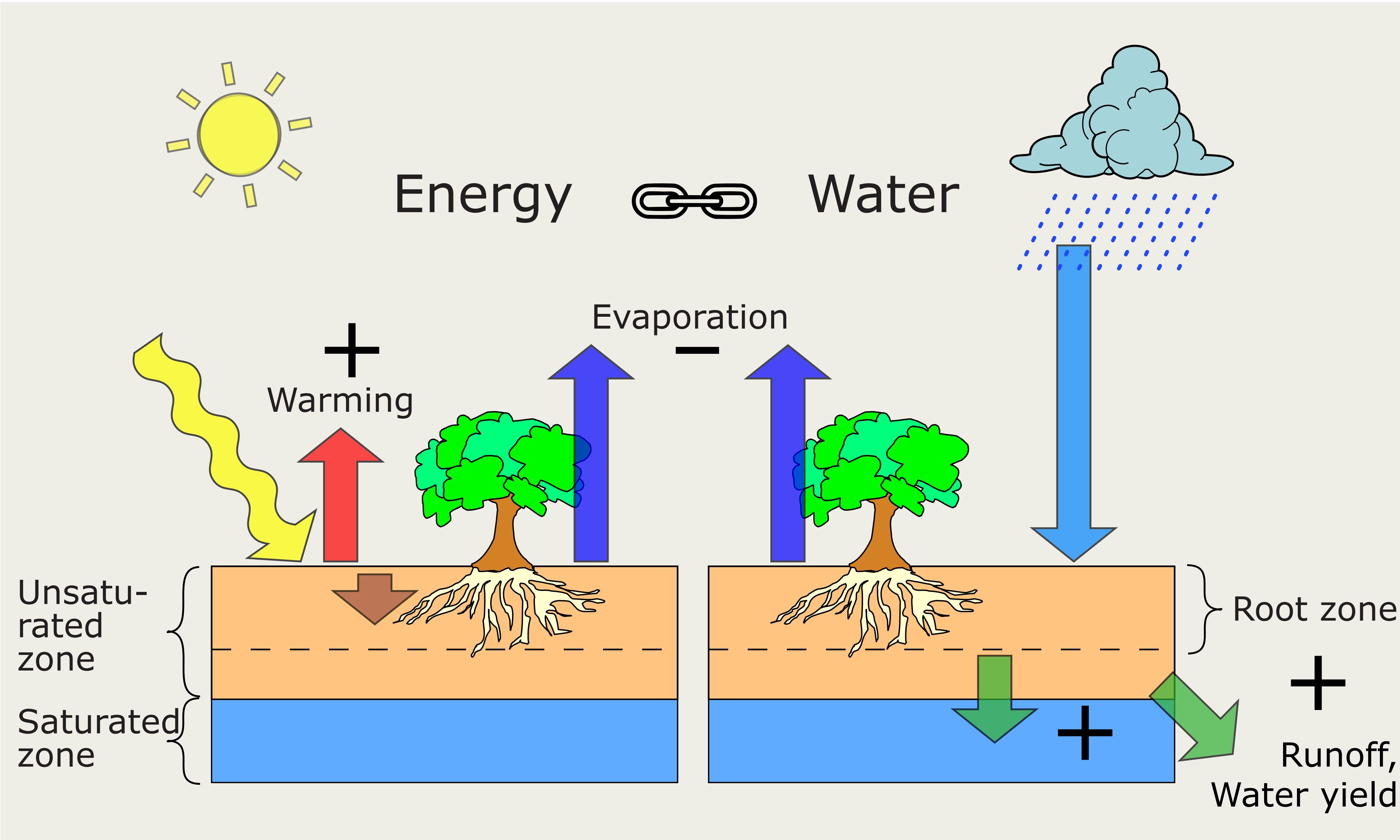


adapted from Sonia Seneviratne

Energy associated with water phase changes



Evaporation, climate, and the hydrological cycle



Modified after Seneviratne et al. (2010), *Earth-Sci. Rev.*, **99**

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Actual versus potential evapotranspiration

Evapotranspiration (ET) is the sum of evaporation and plant transpiration from the Earth's surface to the atmosphere.

Potential evapotranspiration (PET) is a representation of the environmental demand for evapotranspiration and represents the evapotranspiration of "*a short green crop, completely shading the ground, of uniform height and with adequate water status in the soil profile*". It reflects the energy available to evaporate water, and wind available to transport the water vapour from the ground up into the atmosphere. PET can be estimated from routine meteorological observations.

Reference evapotranspiraiton (ET_0) is "the rate of evapotranspiration from a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 sec m^{-1} and an albedo of 0.23"

Actual evapotranspiration is said to equal potential evapotranspiration when there is ample water.

source: wikipedia

Estimating potential evaporation

Penman-Monteith
+ Strongest theoretical basis, advection (vapor pressure deficit)
– Input requirements, need for surface parameters

Priestley Taylor
+ Robust
– Equilibrium ET, no advection

Makkink
+ Global rather than net radiation
– Empirical, derived for Dutch conditions

Thornthwaite
+ Little input data needed
– Does not consider energy balance

Energy **VPD**

$$\lambda ET = \frac{\Delta (R_n - G) + \rho_a c_p \frac{(e_s - e_a)}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)}$$

$$PET = \alpha \frac{\Delta}{\Delta + \gamma}(R_n - G),$$

$$\lambda ET = \alpha_m [s/(s+\gamma)] R_G$$

$$PET = 16\left(\frac{L}{12}\right)\left(\frac{N}{30}\right)\left(\frac{10 T_a}{I}\right)^{\alpha}$$

Observing actual evapotranspiration

Evaporation pans

+ Cheap, robust

– Complex relation to actual ET

Lysimeter

+ Accurate, high-resolution

– Expensive

Catchment water balance

(ET = P – Q)

+ Accurate, sampling volume

– Temporal resolution (dS/dt = 0)

Eddy covariance

+ All ecosystems

– Expensive, footprint definition

Thermal remote sensing

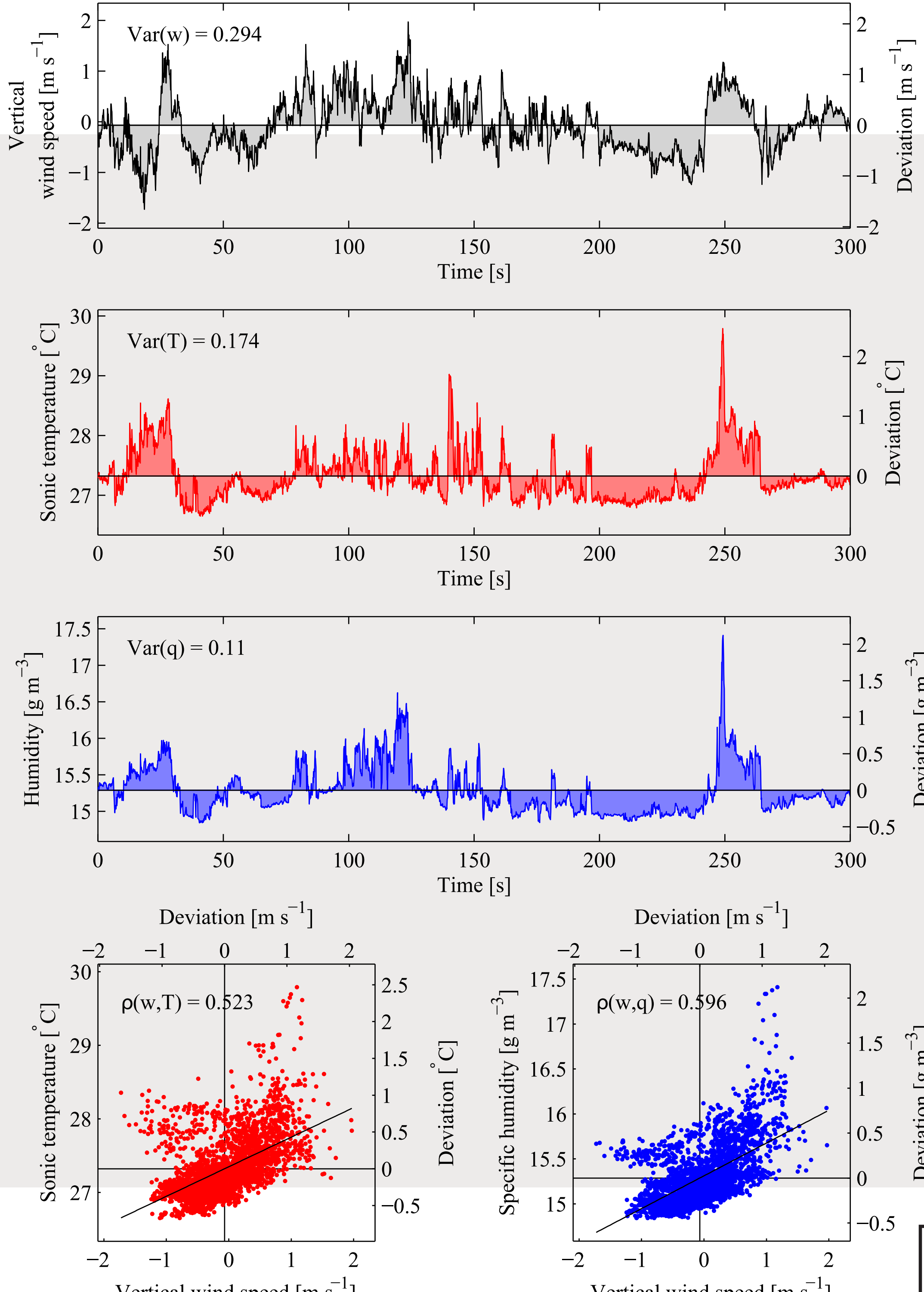
+ Coverage

– Surface temperature not directly related to evaporation

Eddy covariance

Most transport of water vapor occurs via turbulent exchange. Under certain assumptions, the vertical flux is proportional to the covariance between vertical wind speed and the variable of interest (temperature for H and q for ET).

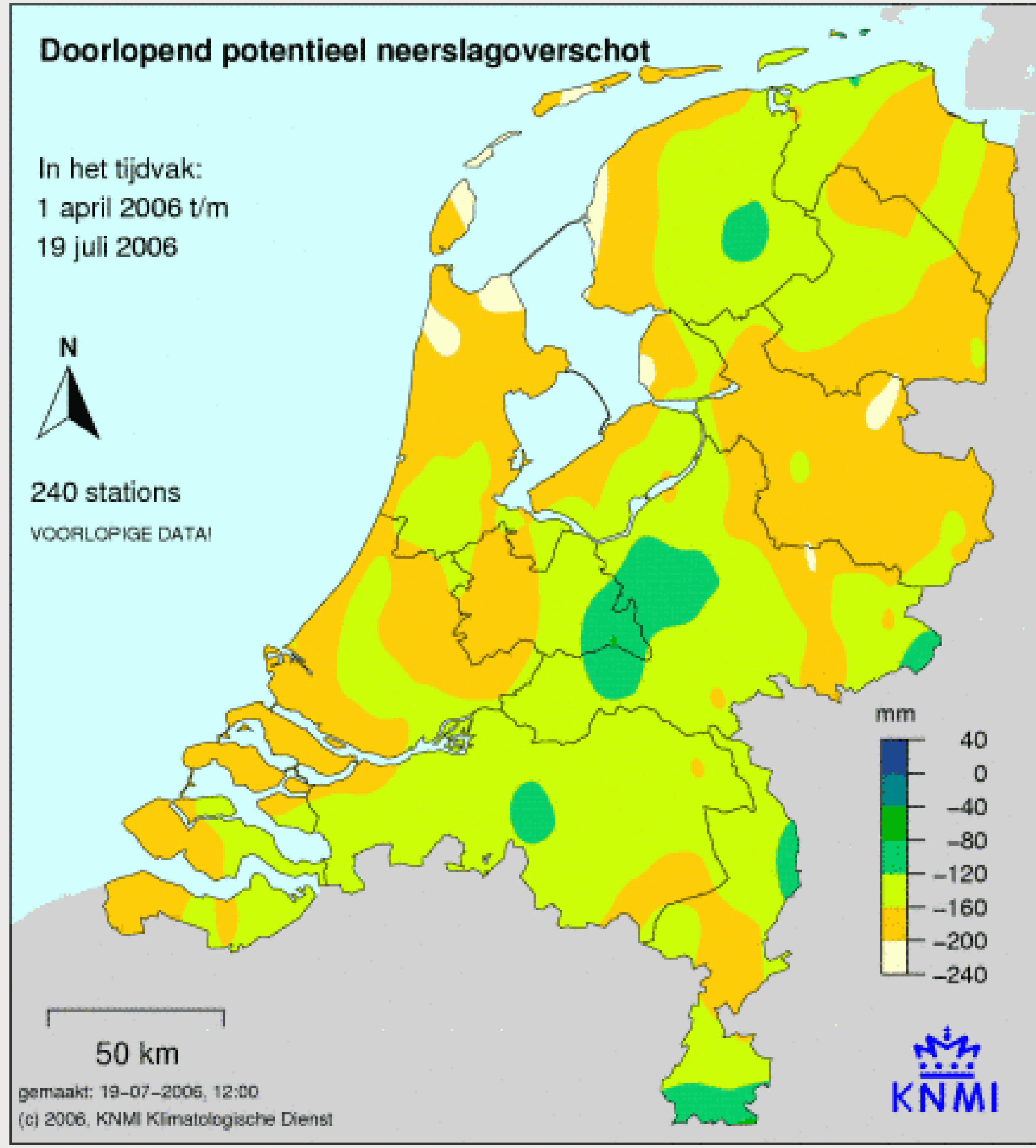
$$F \approx \overline{\rho_a}\,\overline{w'T'}$$



Evaporation monitoring in The Netherlands

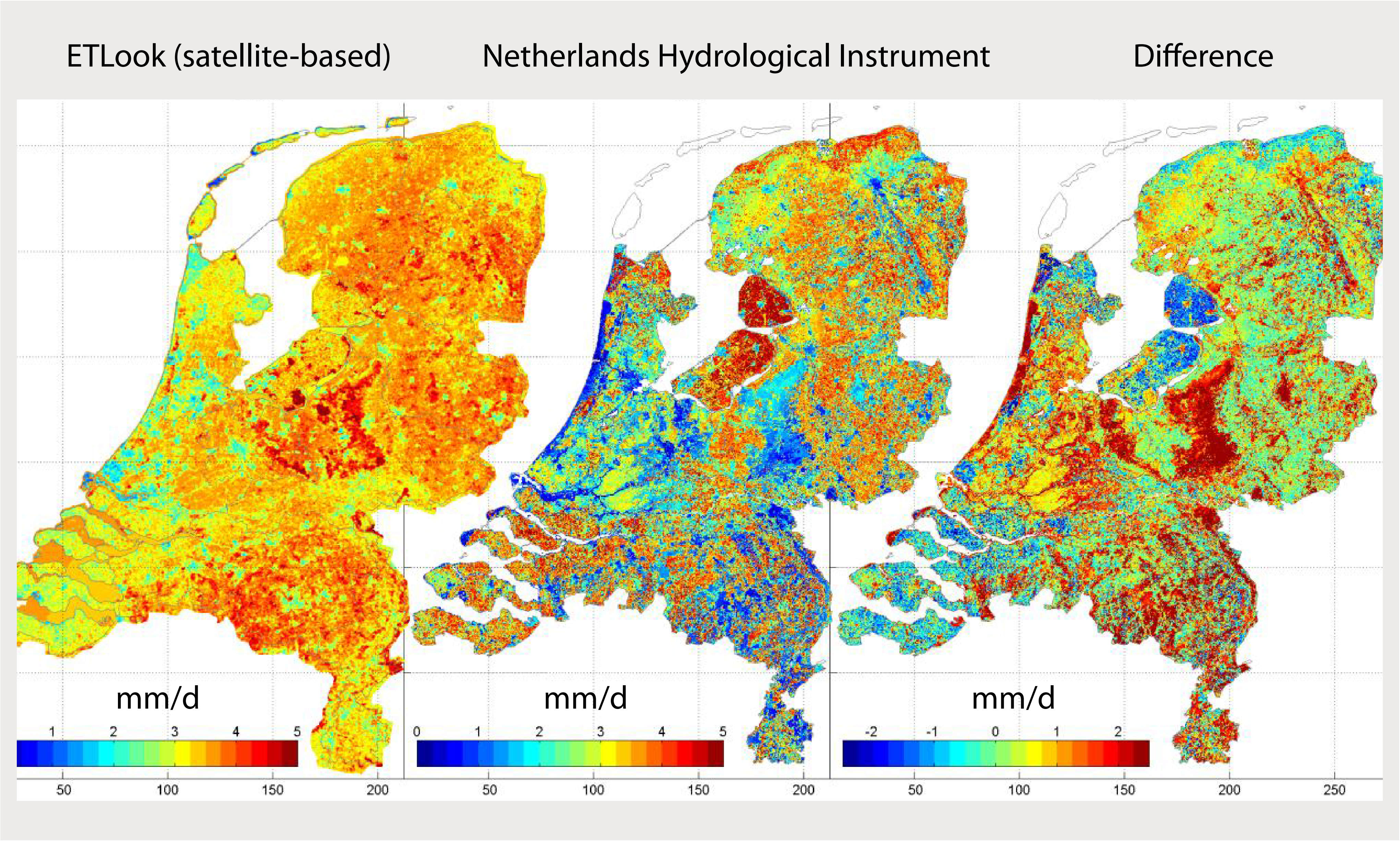
KNMI monitors reference crop evapotranspiration at 25 automatic weather station. Values are used to calculate potential precipitation deficit from 1 April to 31 September. using the Makkink method.

Actual evapotranspiration is being monitored only at selected research sites (e.g. Loobos, Cabauw).



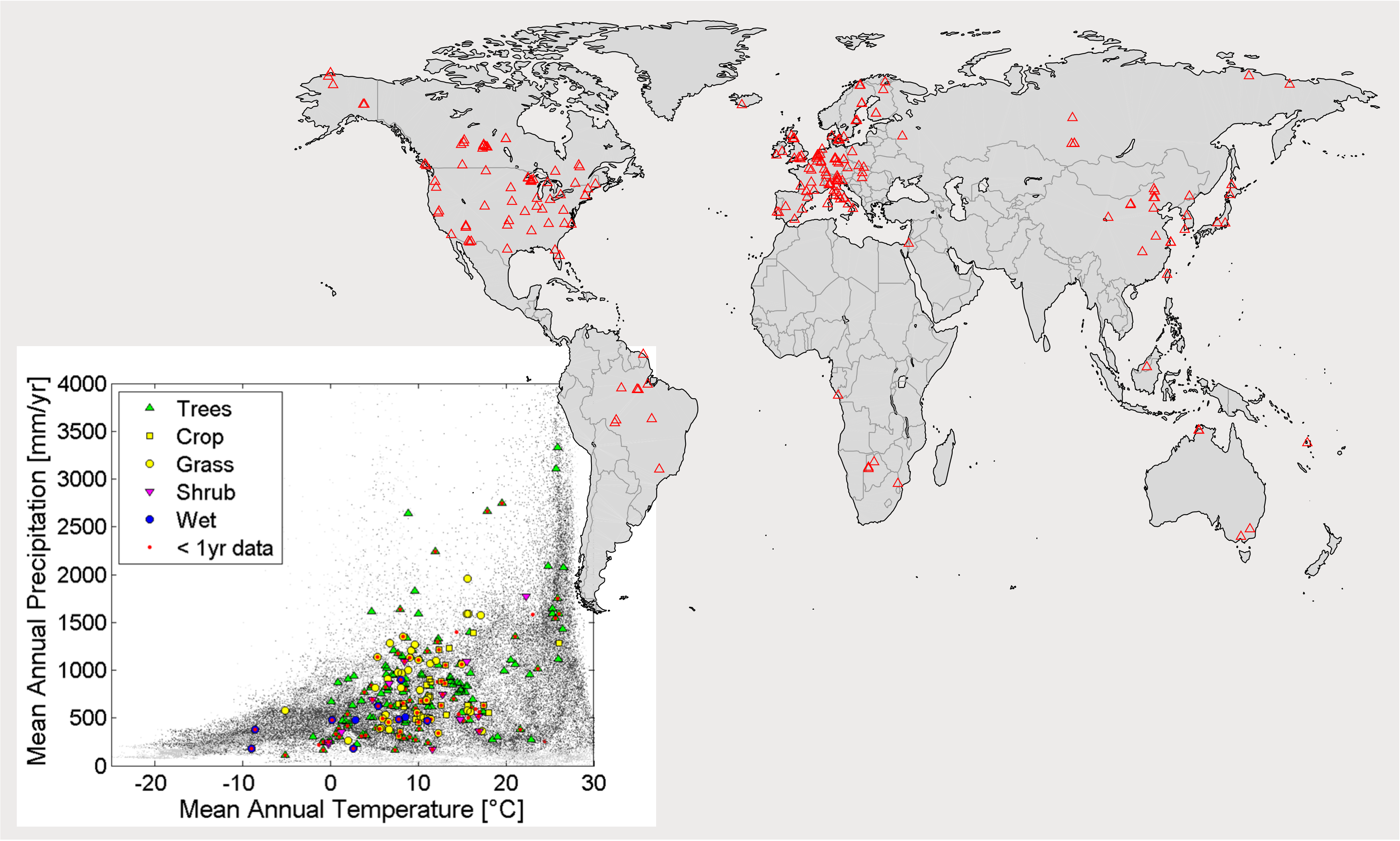
source: knmi.nl

Uncertainty in ET estimates: July 2006 drought



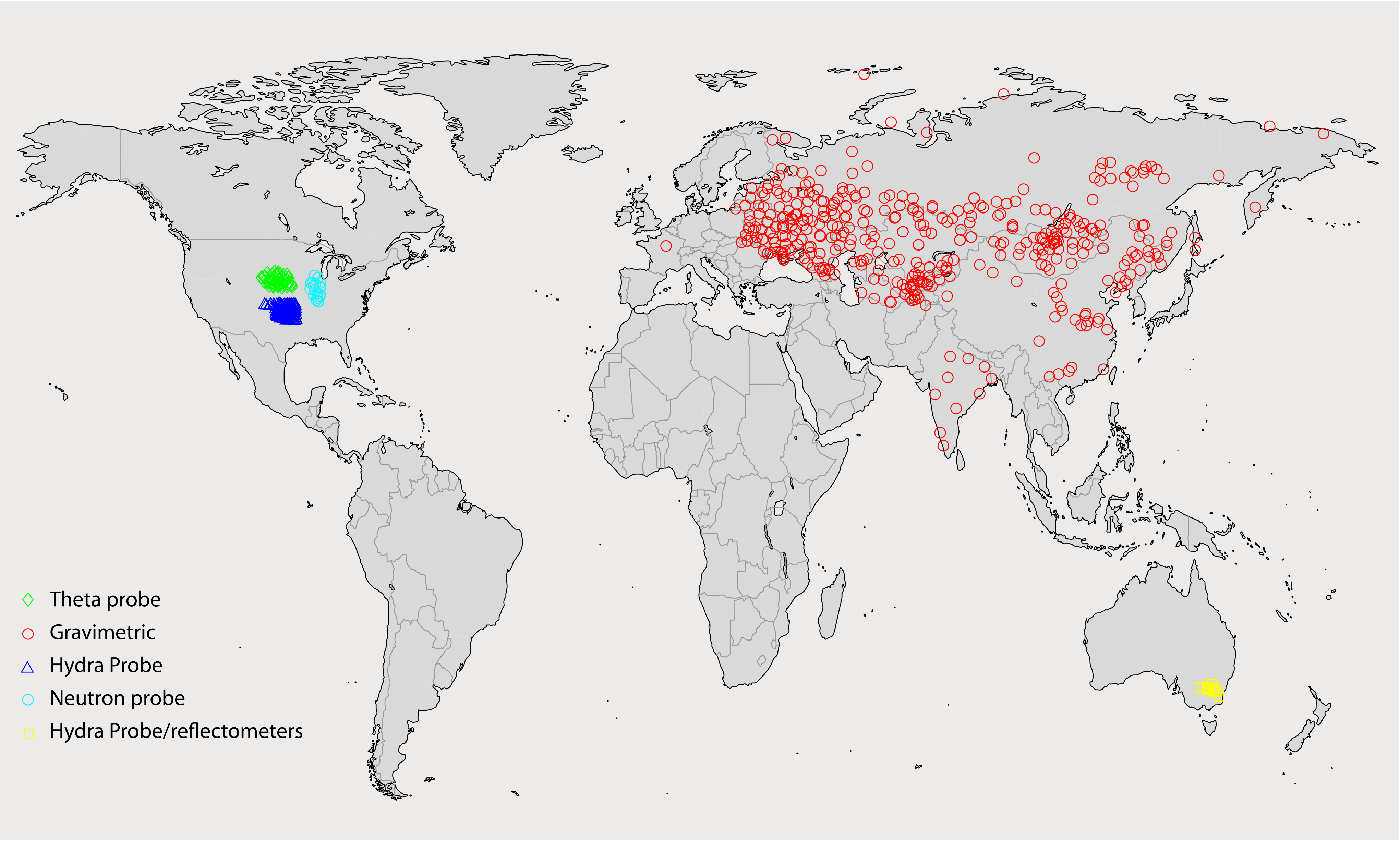
Artesia Water research Unlimited. 2012. Vergelijking van enkele schattingsmethoden voor de actuele verdamping.

Eddy covariance flux towers (FLUXNET)



Baldocchi et al., 2001. FLUXNET: A New Tool to Study the Temporal and Spatial Variability of Ecosystem-Scale Carbon Dioxide, Water Vapor, and Energy Flux Densities**.** *Bull. Am. Meteorol. Soc.* **82**

Long-term soil moisture monitoring networks



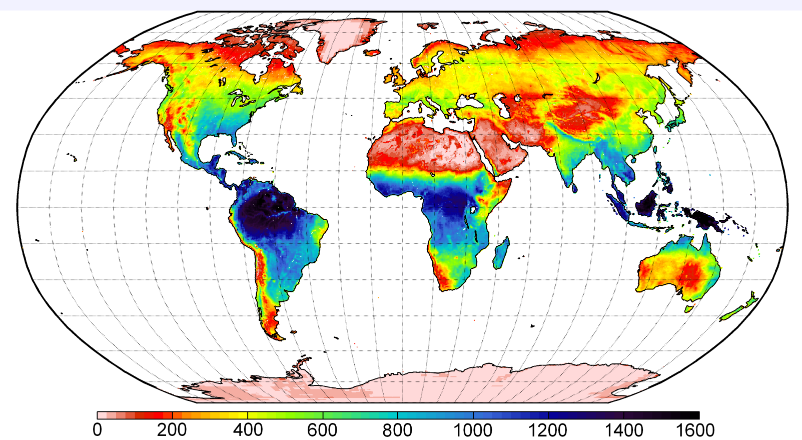
Seneviratne et al., 2010. Investigating soil moisture–climate interactions in a changing climate: A review. *Earth Sci. Rev.* **99**
Robock et al., 2000. The Global Soil Moisture Data Bank**.** *Bull. Am. Meteorol. Soc.* **81**

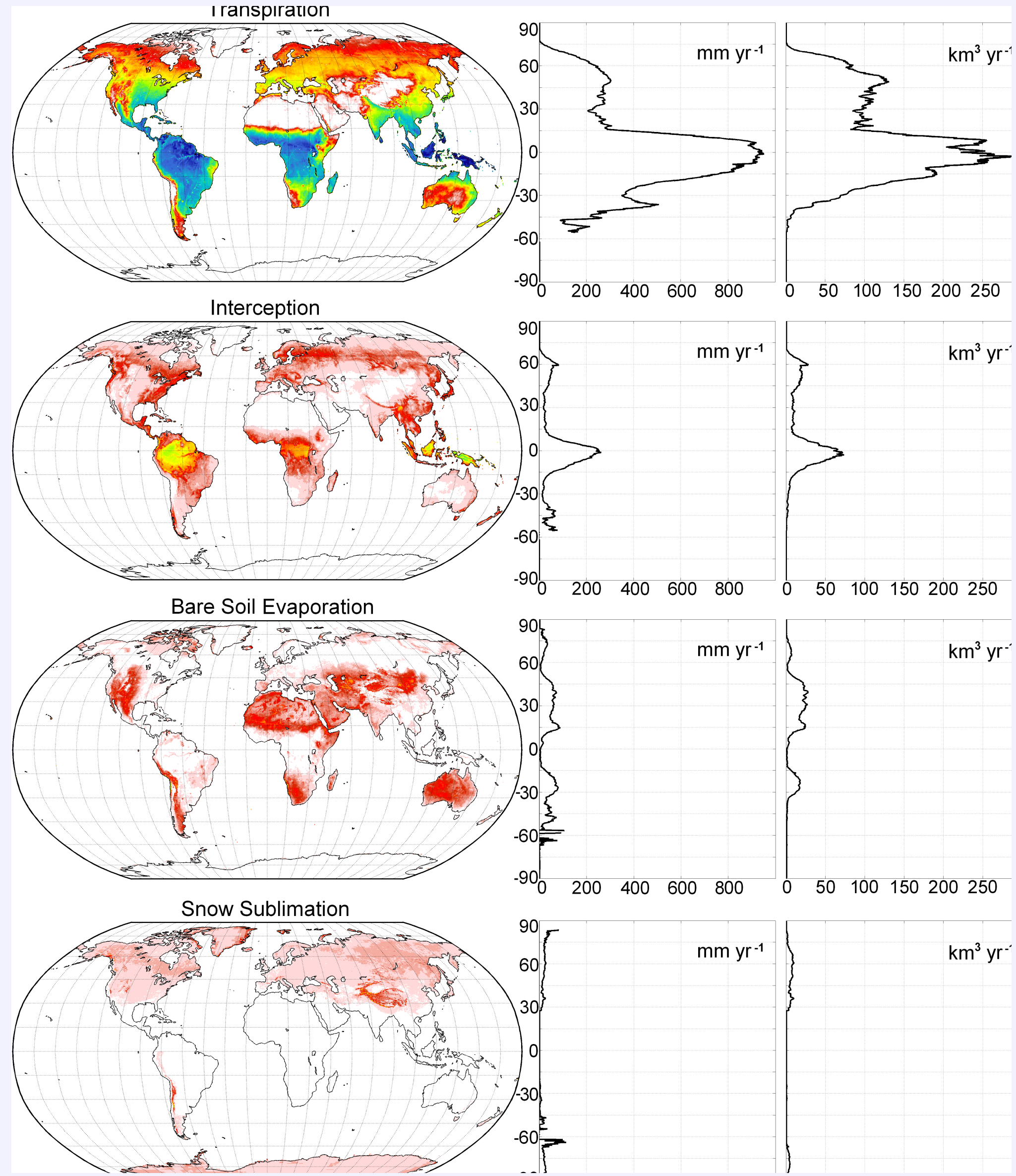
Partitioning of terrestrial evapotranspiration

Globally, most of the evapotranspiration stems from transpiration, which peaks in the tropics.

Interception losses are high in the tropics and northern midlatitudes.

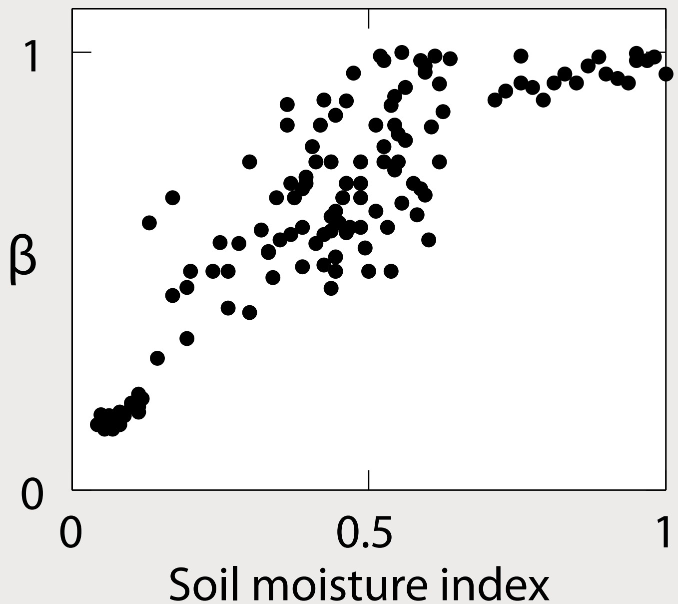
Bare soil evaporation only dominates in (semi-)arid regions





Miralles et al., 2011. Magnitude and variability of land evaporation
and its components at the global scale. *Hydrol. Earth Syst. Sci.* **15**

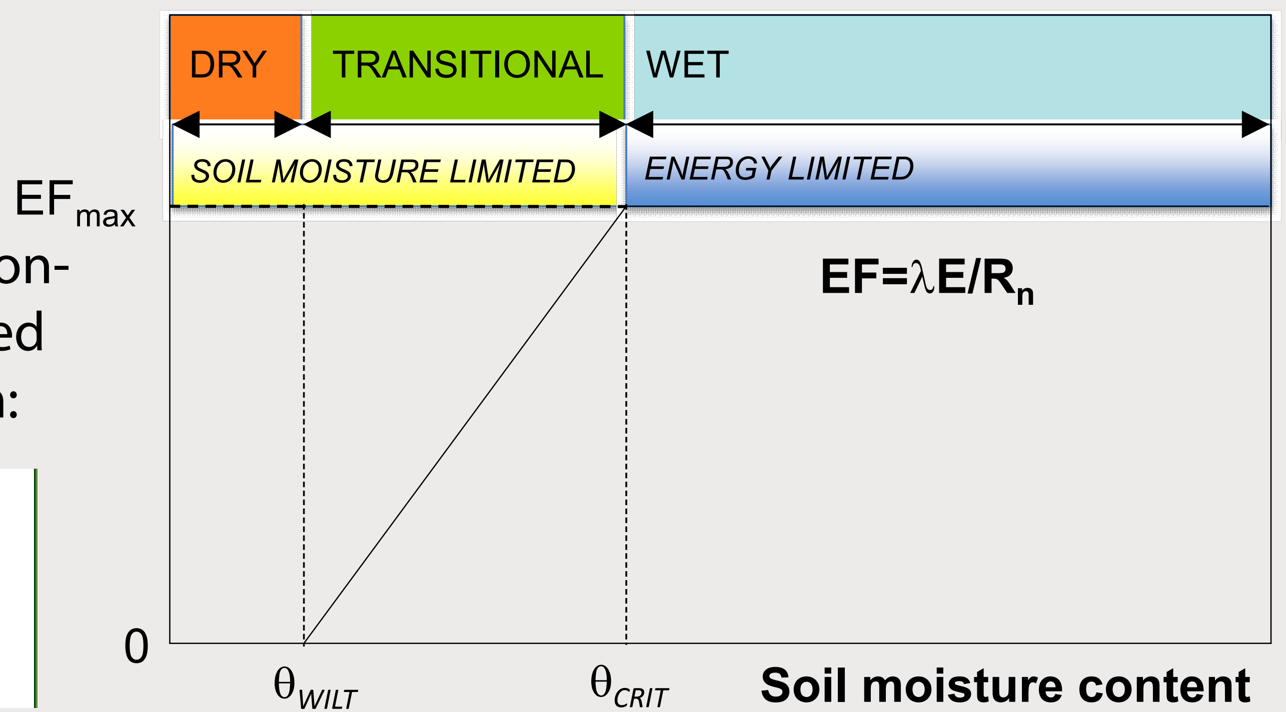
Soil moisture and evaporation



At low soil moisture, uptake from plant roots can no longer satisfy the atmospheric demand (PET) and ET becomes supply or soil moisture limited. This effect is often accounted for by multiplying PET by a soil moisture stress factor β.

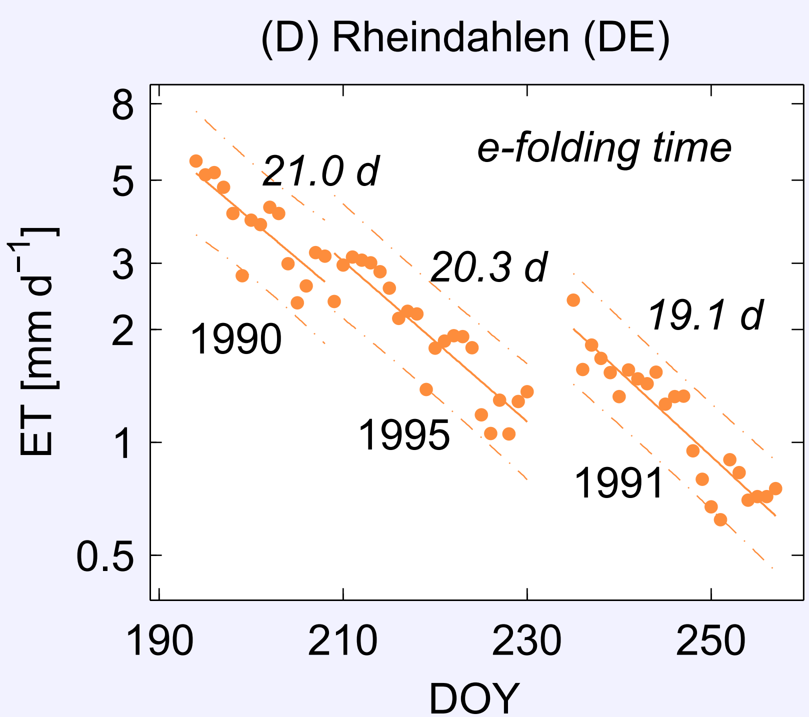
The stress factor is typically non-linear, and often parameterized by a piecewise linear function:

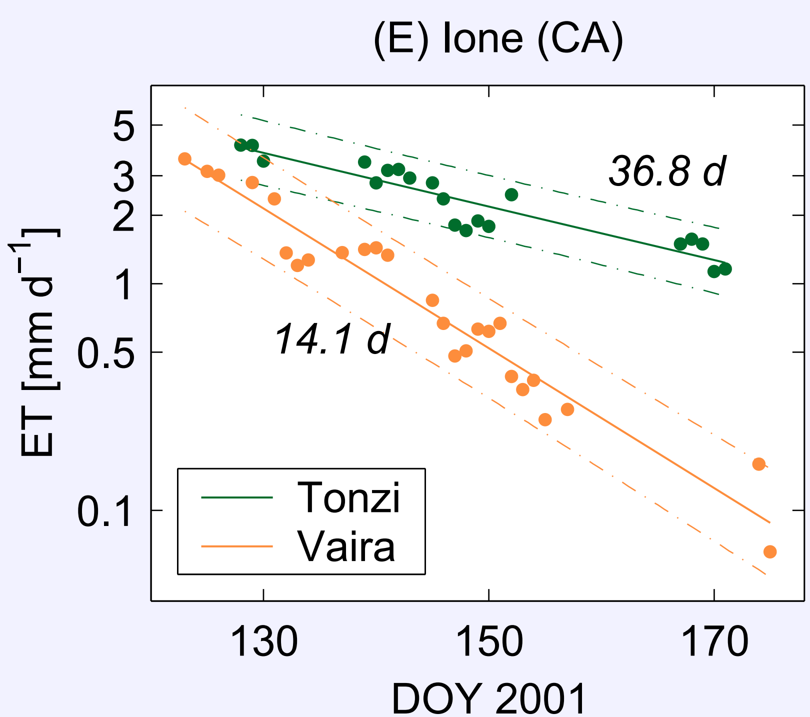
$$\beta(\theta) = \begin{cases} 0, & \theta \leq \theta_w \\ \frac{\theta - \theta_w}{\theta_c - \theta_w}, & \theta_w < \theta \leq \theta_c \\ 1, & \theta_c < \theta \leq \theta_s, \end{cases}$$



Seneviratne et al., 2010. Investigating soil moisture–climate interactions in a changing climate: A review. *Earth Sci. Rev.* **99**
Homaee et al., 2002. Simulation of root water uptake. II. Non-uniform transient water stress using different reduction functions. *Agricul. Wat. Mngt.* **57**

Evapotranspiration: soil moisture impact

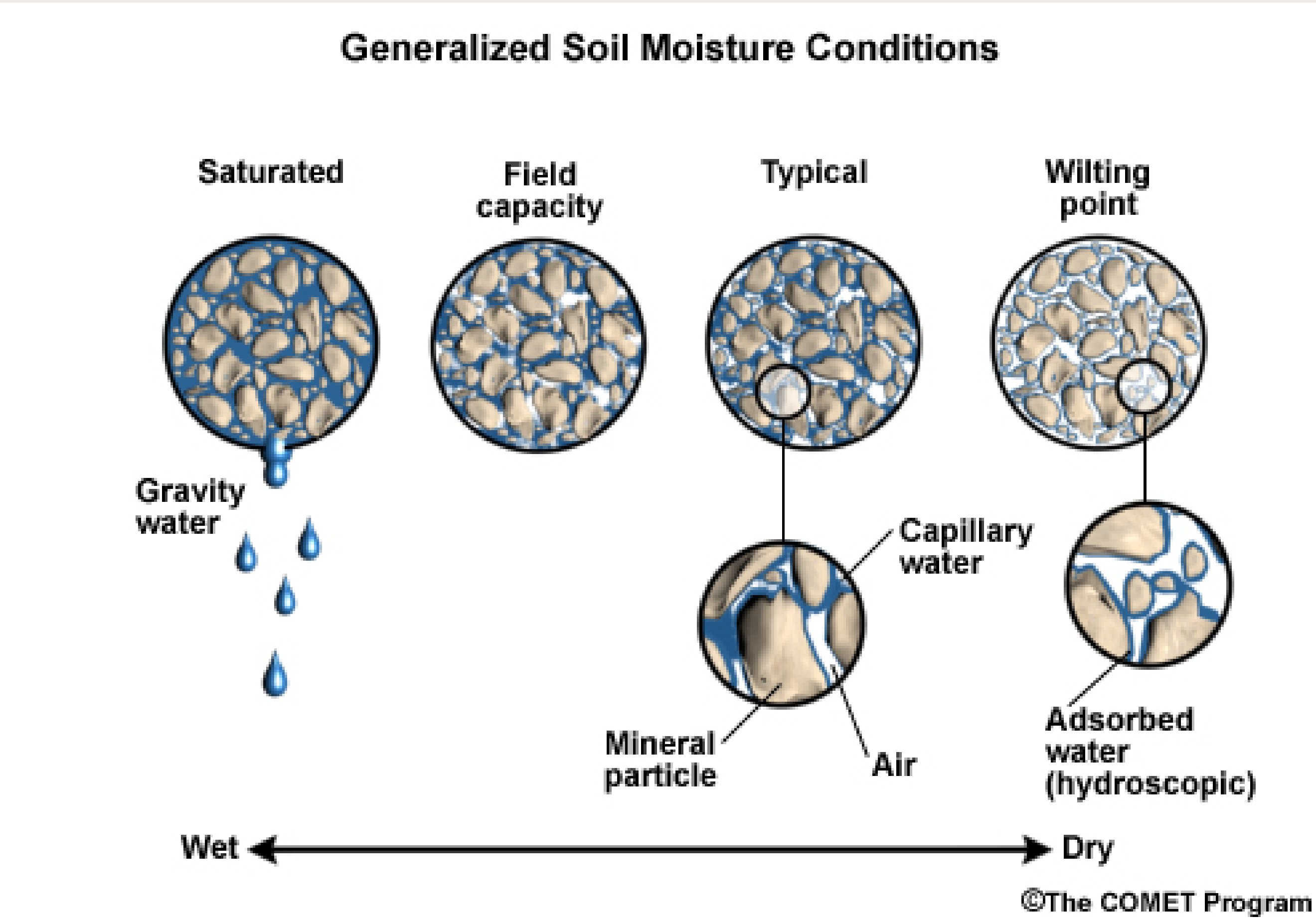




Higher sensitivity of evapotranspiration to soil moisture depletion over grassland than over forest during drought conditions

Teuling et al. (2006), Observed timescales of evapotranspiration response to soil moisture, *Geophys. Res. Lett.* **33**

Soil moisture

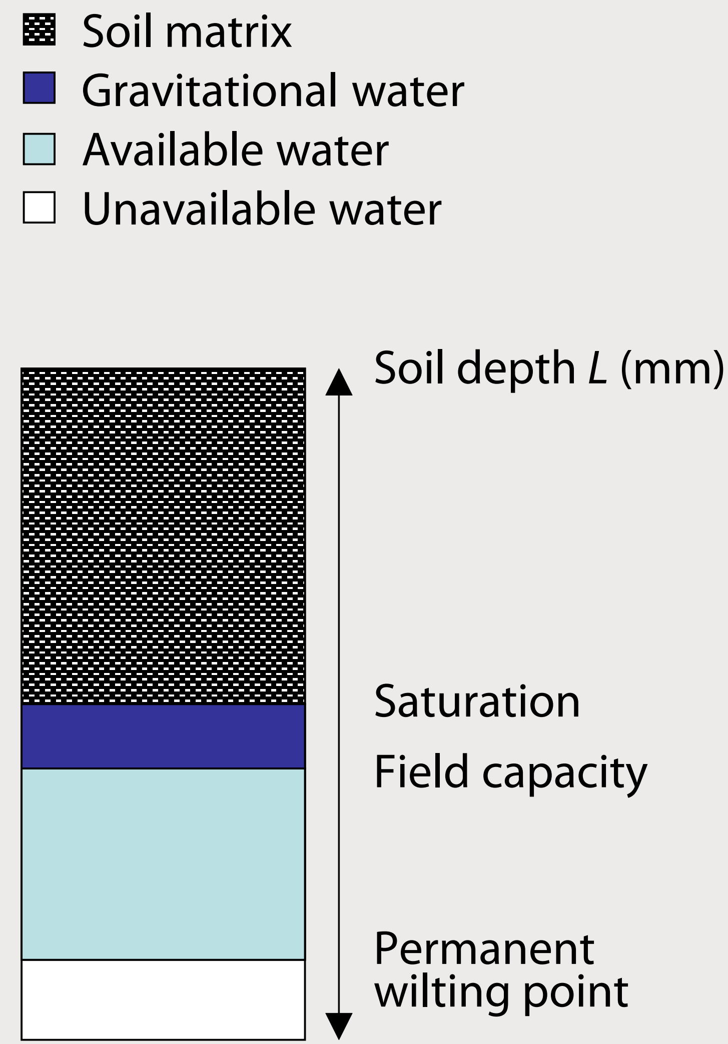


Soil consists of soil particles, air, and water. Soil moisture is the amount of water present in the soil. Due to the small size of the pores, soil moisture is subject to strong capillary forces and large negative pressure. The groundwater level defines the zero- or atmospheric pressure plane; below it the pore pressure is positive and the soil is fully saturated.

http://www.comet.ucar.edu/

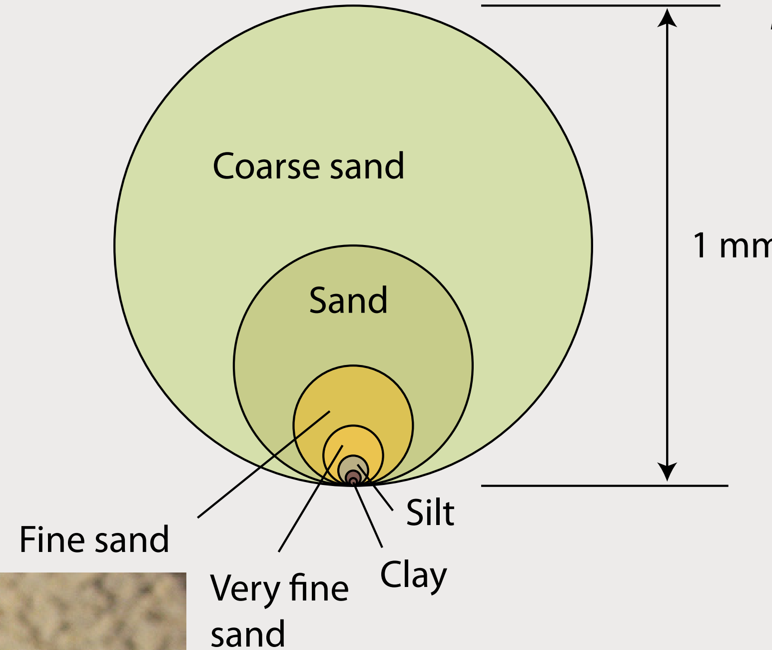
Characteristic soil moisture levels and units

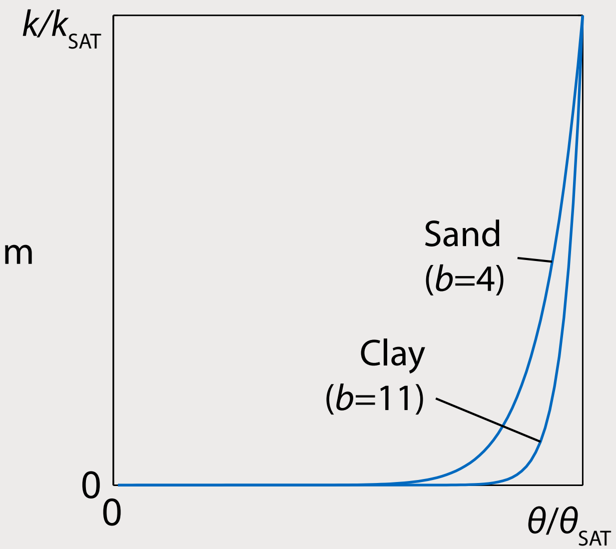
Absolute water content [mm] $S = \theta \cdot L$	Ratio of saturation or wetness [mm/mm] $s = \theta/\theta_{SAT}$	Volumetric water content [mm/mm] θ	Soil moisture potential [m] or [10^2 hPa] h or Ψ	Soil moisture index [-] *SMI*
		1		
$\theta_{SAT} \cdot L$	1	θ_{SAT}	−0.01	
		θ_{FC}	−1–−3	1
		θ_{CRIT}		
		θ_{WILT}	−150	0
0	0	0		



Seneviratne et al., 2010. Investigating soil moisture–climate interactions in a changing climate: A review. *Earth Sci. Rev.* **99**

Soil type and hydraulic conductivity





Relation between volumetric water content θ and unsaturated conductivity k can be described by a power-law:

$$k(\theta) = k_{SAT}\,(\theta/\theta_{SAT})^{2b+3}$$

where k_{SAT} is the saturated hydraulic conductivity at full saturation (porosity).

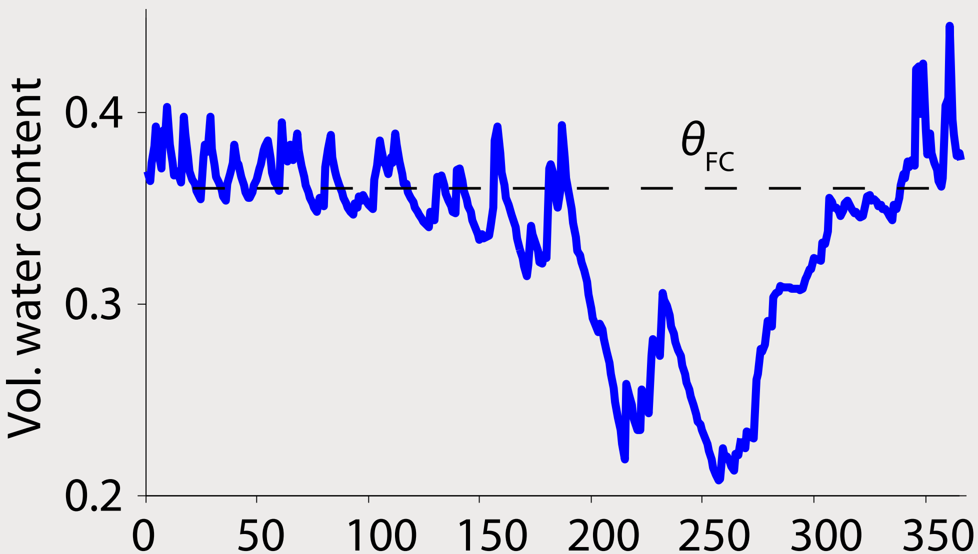
Representative values for soil hydraulic properties

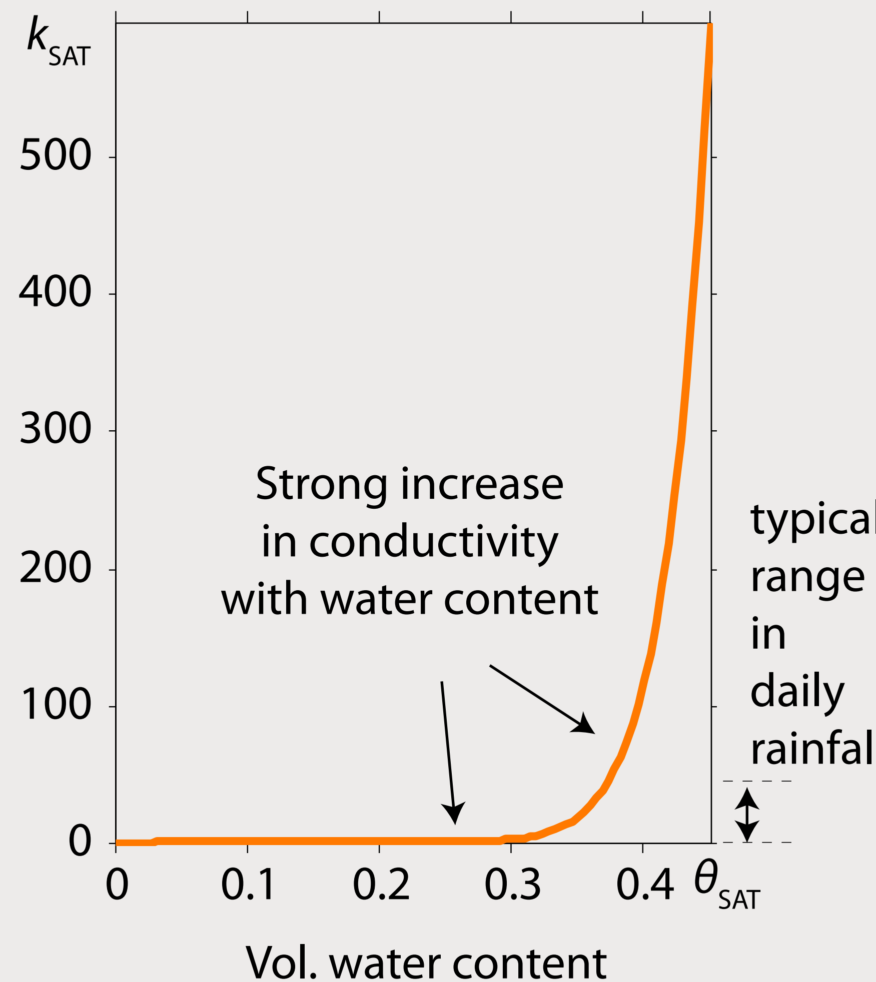
Soil texture	# Soils	Mean clay fraction	b	$\theta_{SAT} \pm \sigma$	k_{SAT} [m/d]
Sand	13	0.03	4.05	0.395 ± 0.056	15.21
Loamy sand	30	0.06	4.38	0.410 ± 0.068	13.51
Sandy loam	204	0.09	4.90	0.435 ± 0.086	3.00
Silt loam	384	0.14	5.30	0.485 ± 0.059	0.62
Loam	125	0.19	5.39	0.451 ± 0.078	0.60
Sandy clay loam	80	0.28	7.12	0.420 ± 0.059	0.54
Silty clay loam	147	0.34	7.75	0.477 ± 0.057	0.15
Clay loam	262	0.34	8.52	0.476 ± 0.053	0.21
Sandy clay	19	0.43	10.4	0.426 ± 0.057	0.19
Silty clay	441	0.49	10.4	0.492 ± 0.064	0.09
Clay	140	0.63	11.4	0.482 ± 0.050	0.11

Clapp and Hornberger, 1978. Empirical equations for some soil hydraulic properties. *Water Resour. Res.* **14**

Field capacity

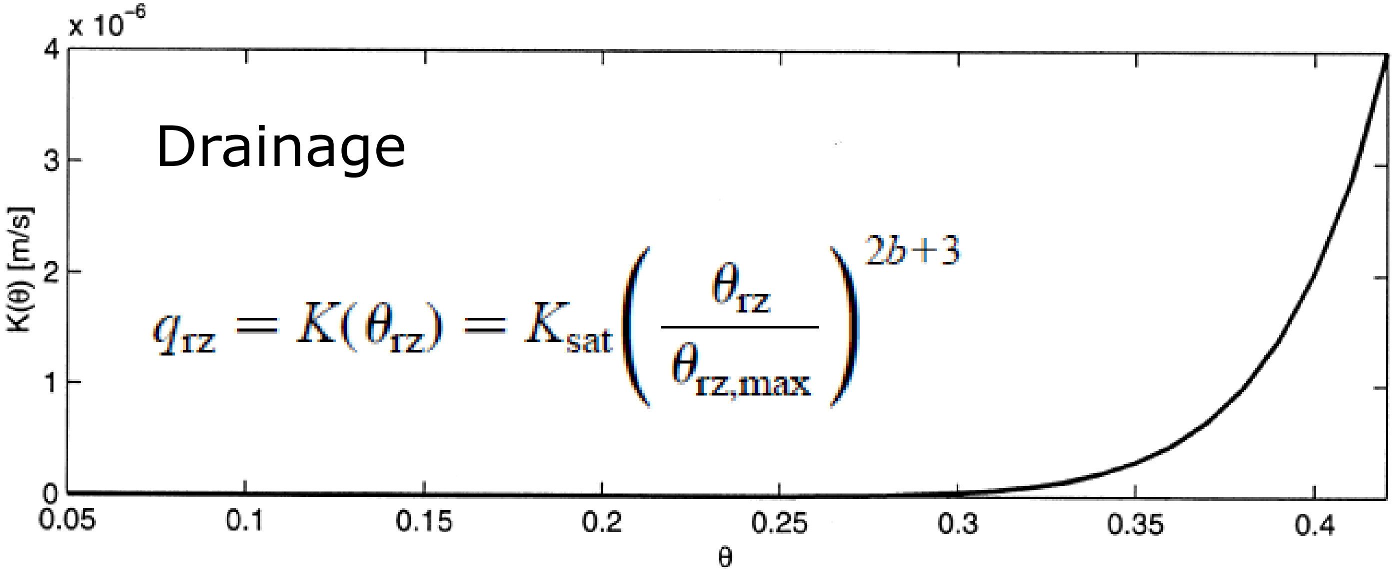
Field capacity (θ_{FC}) is the amount of soil moisture held in the soil after excess water has drained away and the rate of downward movement has decreased. It can be defined as the VWC at a potential of −33 kPa.





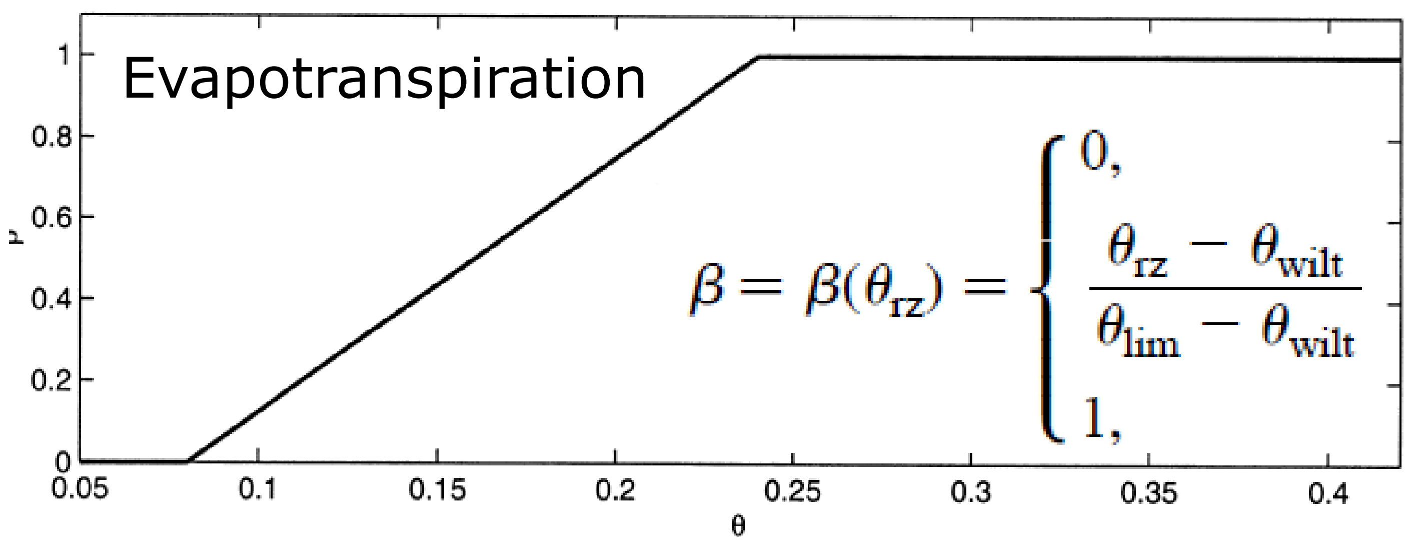
Combined soil moisture losses

Drainage



$$q_{rz} = K(\theta_{rz}) = K_{sat}\left(\frac{\theta_{rz}}{\theta_{rz,max}}\right)^{2b+3}$$

Evapotranspiration

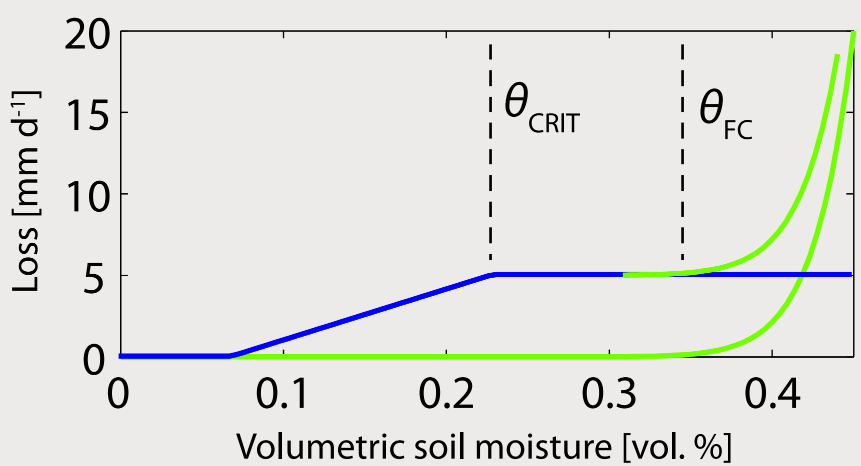


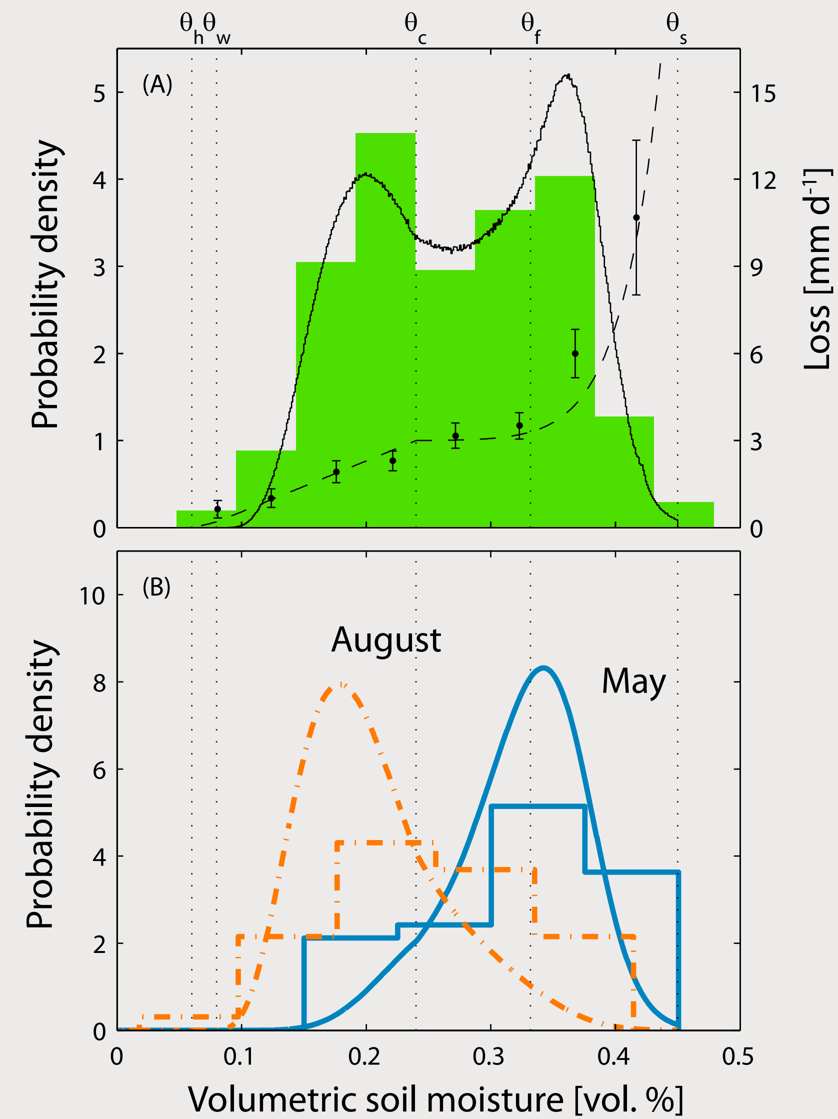
$$\beta = \beta(\theta_{rz}) = \begin{cases} 0, & \text{for } \theta_{rz} \le \theta_{wilt} \\ \dfrac{\theta_{rz} - \theta_{wilt}}{\theta_{lim} - \theta_{wilt}} & \text{for } \theta_{wilt} < \theta_{rz} < \theta_{lim} \\ 1, & \text{for } \theta_{rz} \ge \theta_{lim} \end{cases}$$

Albertson & Kiely (2001), *Journal of Hydrology* **243**

The soil moisture loss function and bimodality

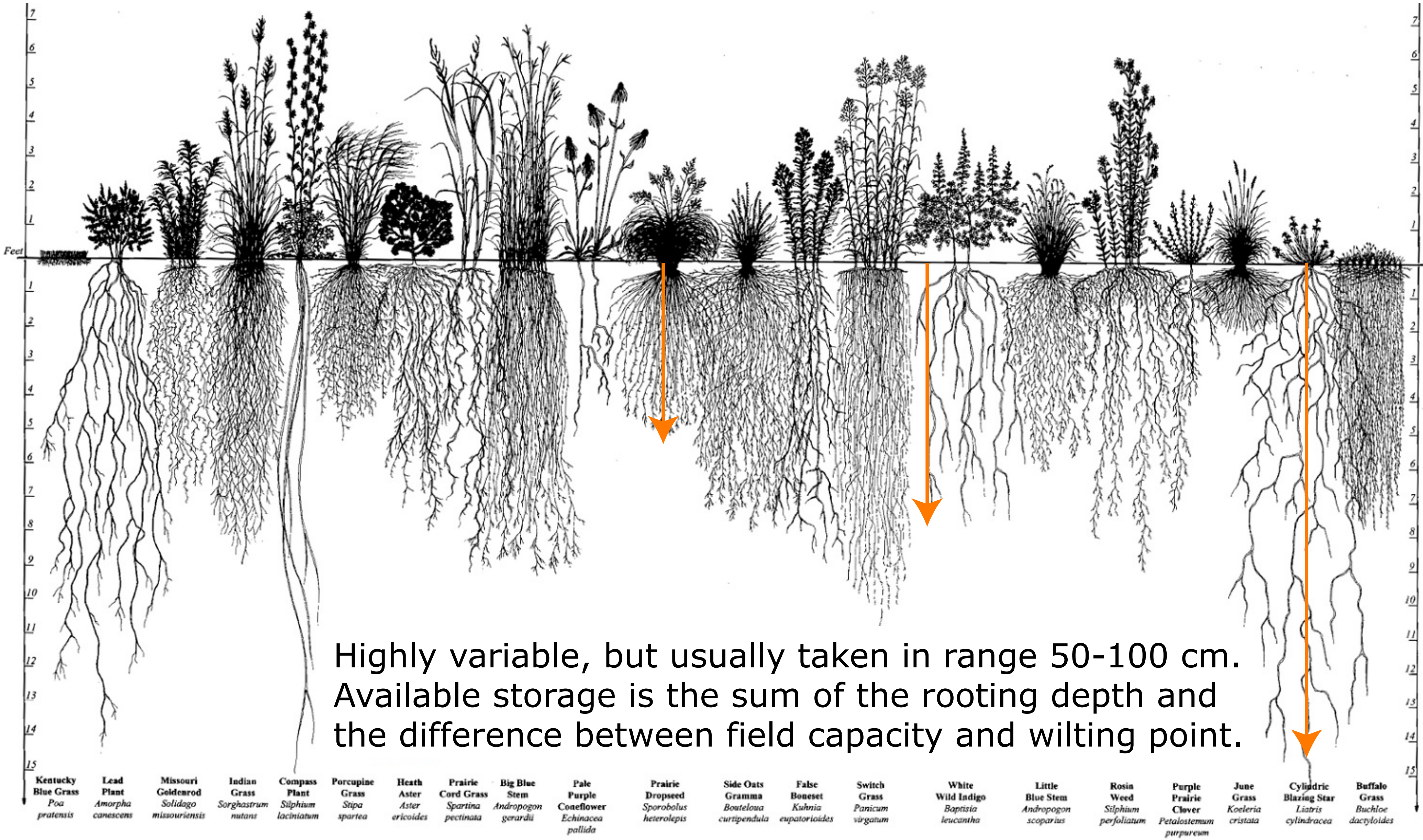
Upward evapotranspiration losses and downward drainage losses combine into the soil moisture loss function. The shape is often strongly nonlinear, with a middle section without sensitivity of losses to soil moisture. As a result, there is a preference for dry or wet soil moisture conditions.





Teuling et al., 2005. On bimodality in warm season soil moisture observations. *Geophys. Res. Lett.* **32**

Rooting depth and available storage



Highly variable, but usually taken in range 50-100 cm. Available storage is the sum of the rooting depth and the difference between field capacity and wilting point.

illustration: Conservation Research Institute

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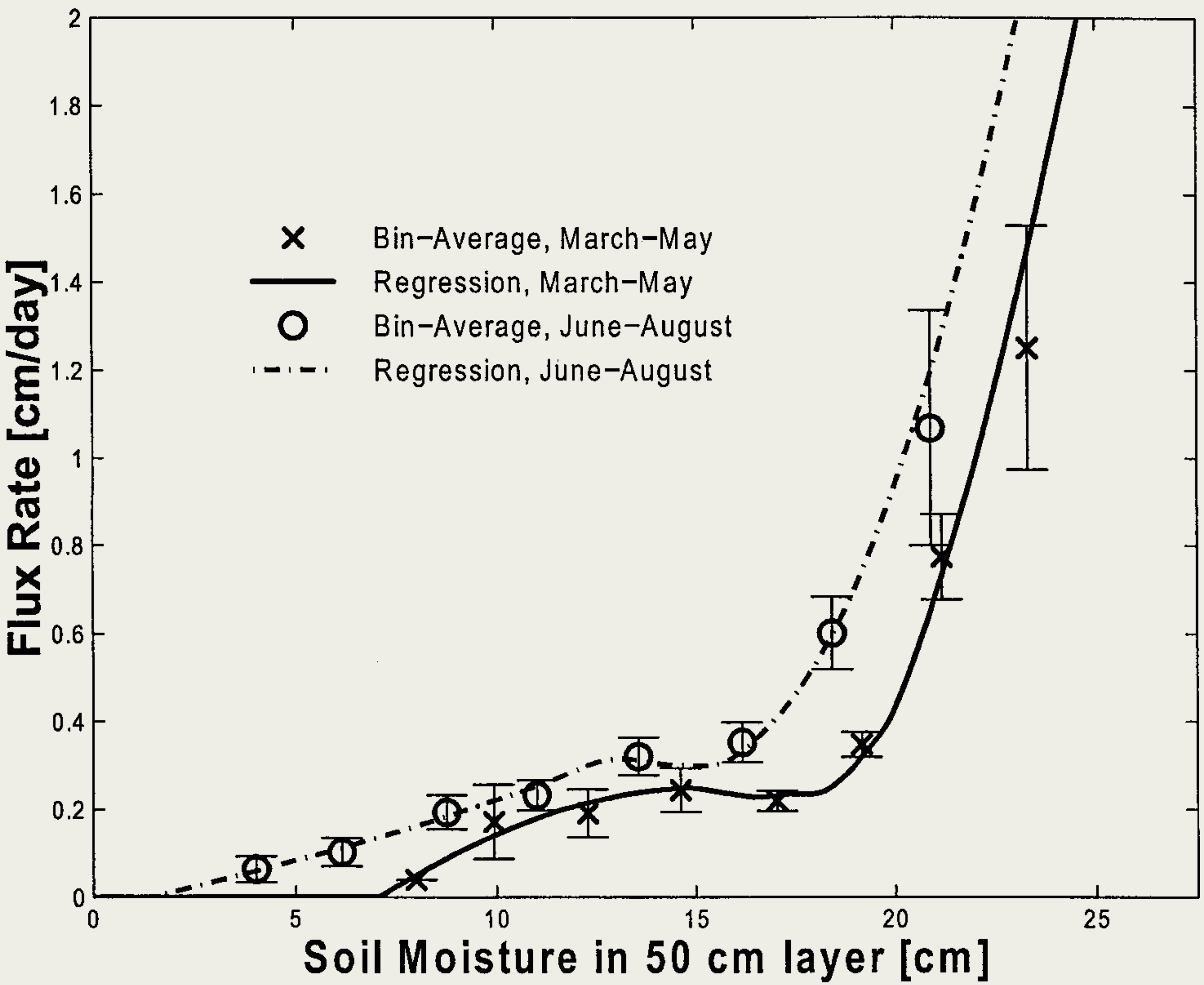
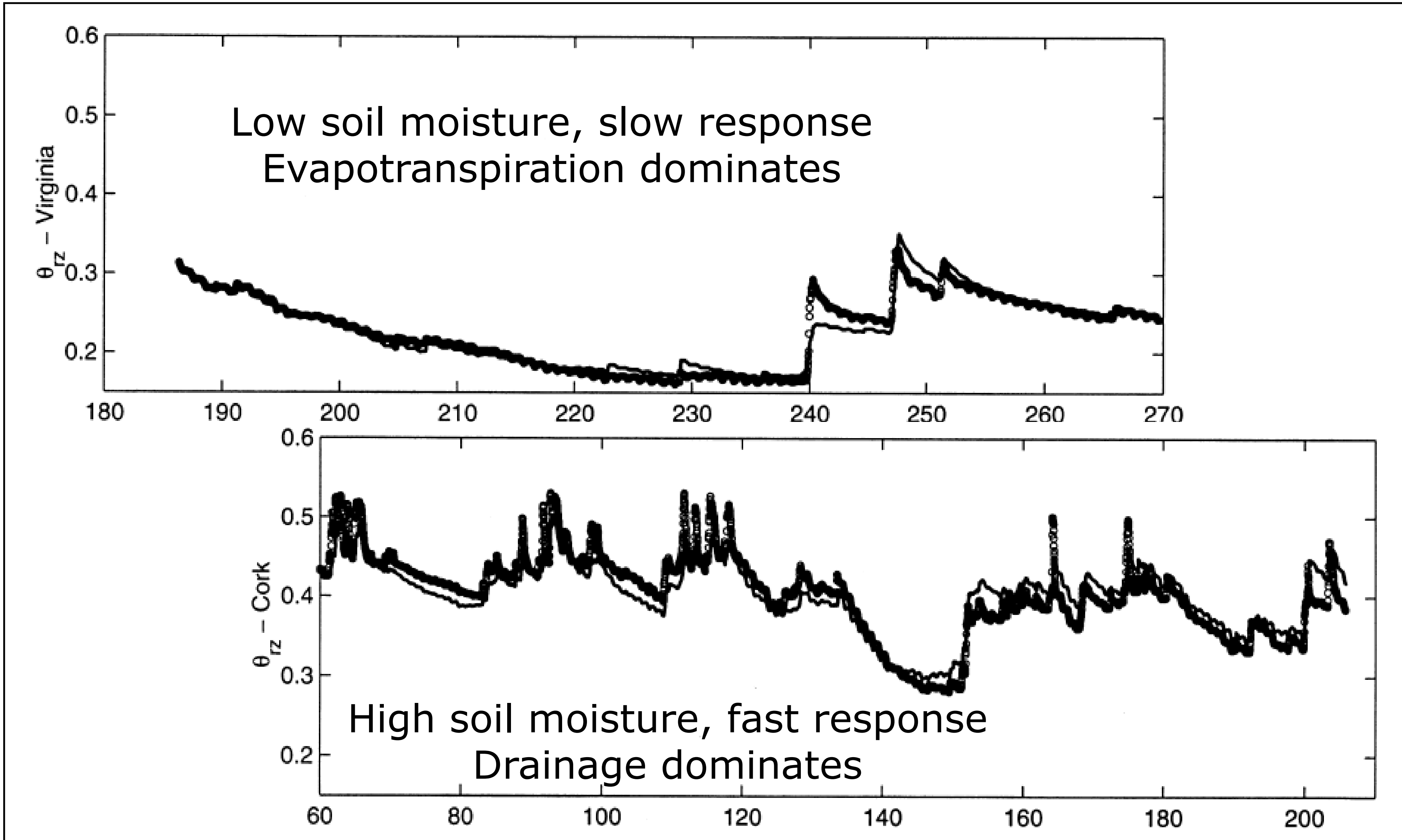
Loss function from observations



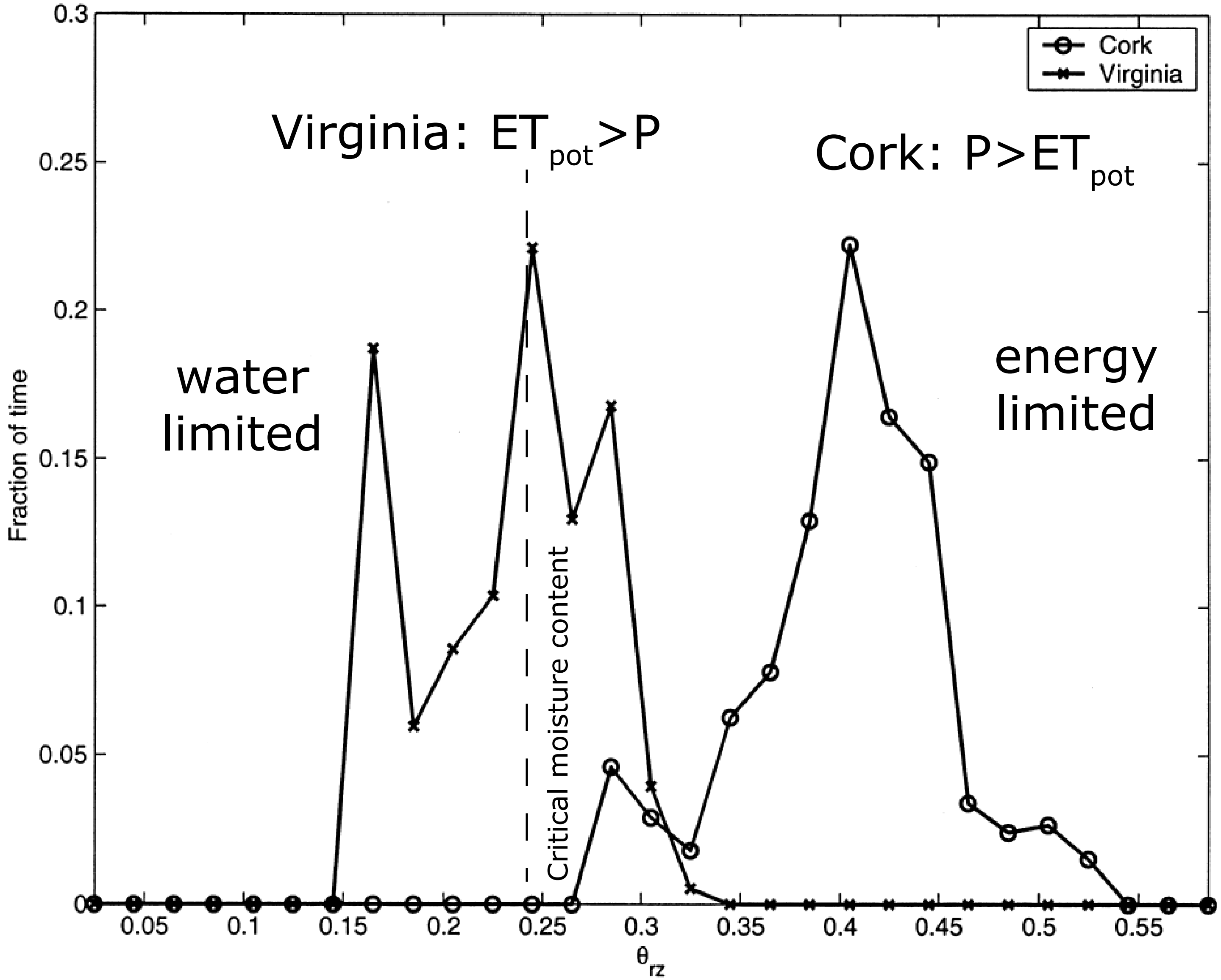
Salvucci (2001), *Water Resour. Res.*, **37**

Contrasting dynamics



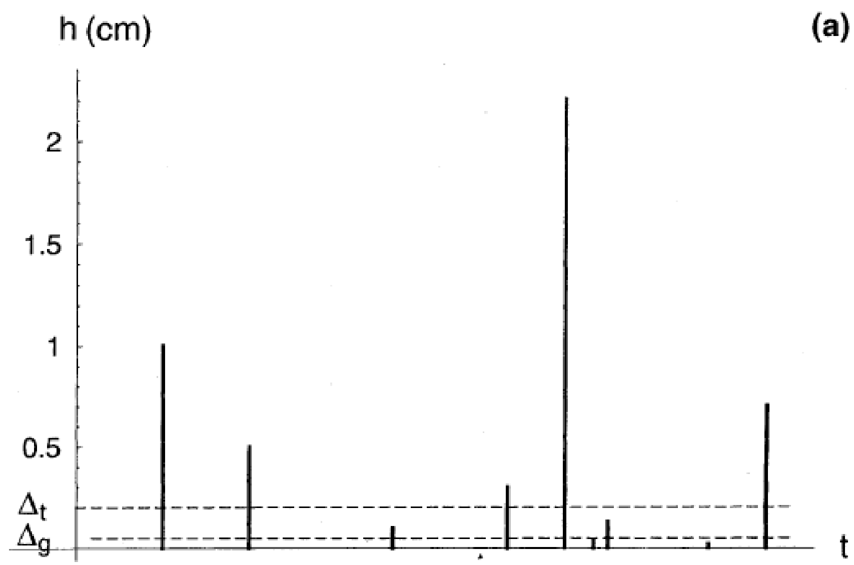
Albertson & Kiely (2001), *Journal of Hydrology* **243**

Different distributions



Albertson & Kiely (2001), *Journal of Hydrology* **243**

Stochastic rainfall



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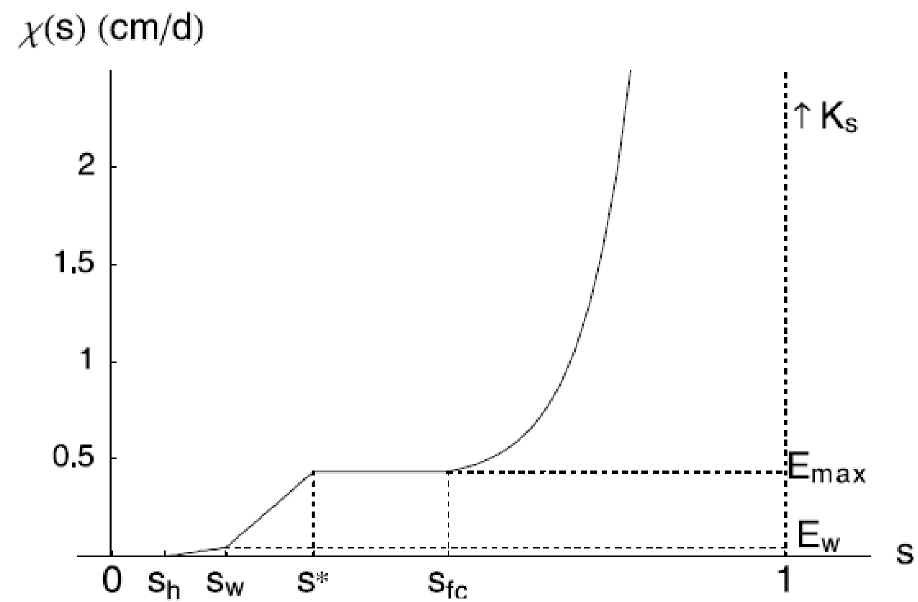

The distribution of the times τ between precipitation events is exponential with mean $1/\lambda$ (e.g., [13]), i.e.,

$$f_{\mathrm{T}}(\tau) = \lambda \mathrm{e}^{-\lambda\tau} \quad \text{for } \tau \geqslant 0. \tag{4}$$

The depth of rainfall events is assumed to be an independent random variable h, described by an exponential probability density function

$$f_{\mathrm{H}}(h) = \frac{1}{\alpha}\mathrm{e}^{-\frac{1}{\alpha}h} \quad \text{for } h \geqslant 0, \tag{5}$$

where α is the mean depth of rainfall events. Since the

=

$$p(s) = \begin{cases} \frac{C}{\eta_{\mathrm{w}}}\left(\frac{s-s_{\mathrm{h}}}{s_{\mathrm{w}}-s_{\mathrm{h}}}\right)^{\frac{\lambda'(s_{\mathrm{w}}-s_{\mathrm{h}})}{\eta_{\mathrm{w}}}-1}\mathrm{e}^{-\gamma s}, \\ \quad s_{\mathrm{h}} < s \leqslant s_{\mathrm{w}}, \\ \frac{C}{\eta_{\mathrm{w}}}\left[1+\left(\frac{\eta}{\eta_{\mathrm{w}}}-1\right)\left(\frac{s-s_{\mathrm{w}}}{s^*-s_{\mathrm{w}}}\right)\right]^{\frac{\lambda'(s^*-s_{\mathrm{w}})}{\eta-\eta_{\mathrm{w}}}-1}\mathrm{e}^{-\gamma s}, \\ \quad s_{\mathrm{w}} < s \leqslant s^*, \\ \frac{C}{\eta}\mathrm{e}^{-\gamma s+\frac{\lambda'}{\eta}(s-s^*)}\left(\frac{\eta}{\eta_{\mathrm{w}}}\right)^{\lambda'\frac{s^*-s_{\mathrm{w}}}{\eta-\eta_{\mathrm{w}}}}, \\ \quad s^* < s \leqslant s_{\mathrm{fc}}, \\ \frac{C}{\eta}\mathrm{e}^{-(\beta+\gamma)s+\beta s_{\mathrm{fc}}}\left(\frac{\eta\,\mathrm{e}^{\beta s}}{(\eta-m)\mathrm{e}^{\beta s_{\mathrm{fc}}}+m\mathrm{e}^{\beta s}}\right)^{\frac{\lambda'}{\beta(\eta-m)}+1} \\ \quad \times\left(\frac{\eta}{\eta_{\mathrm{w}}}\right)^{\lambda'\frac{s^*-s_{\mathrm{w}}}{\eta-\eta_{\mathrm{w}}}}\mathrm{e}^{\frac{\lambda'}{\eta}(s_{\mathrm{fc}}-s^*)}, \\ \quad s_{\mathrm{fc}} < s \leqslant 1. \end{cases}$$

Laio et al. (2001), *Advances in Water Resources* **24**

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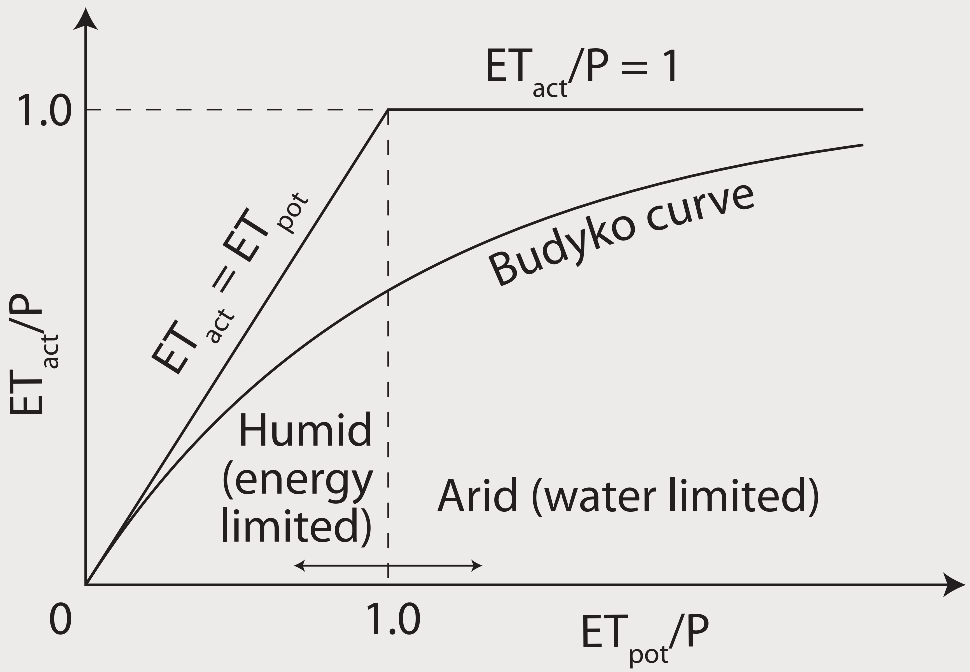
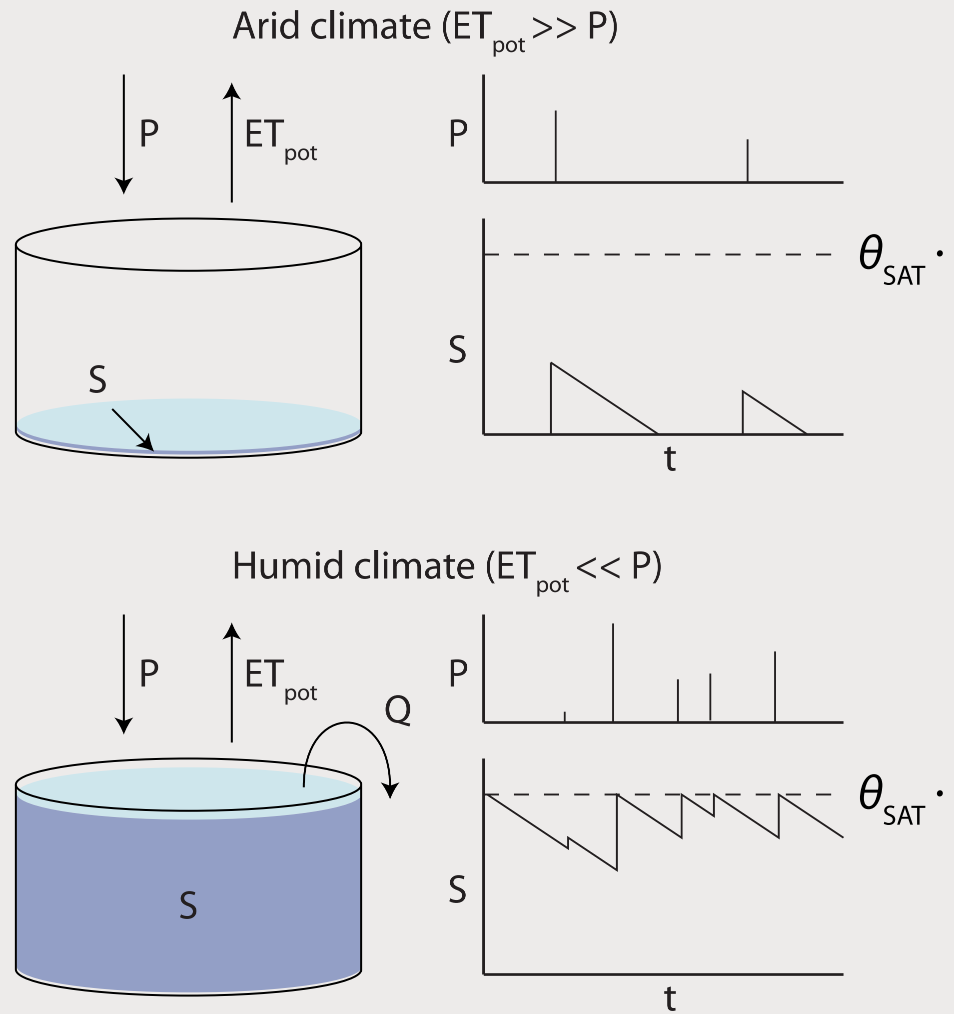
Dry and wet limits to catchment water balance





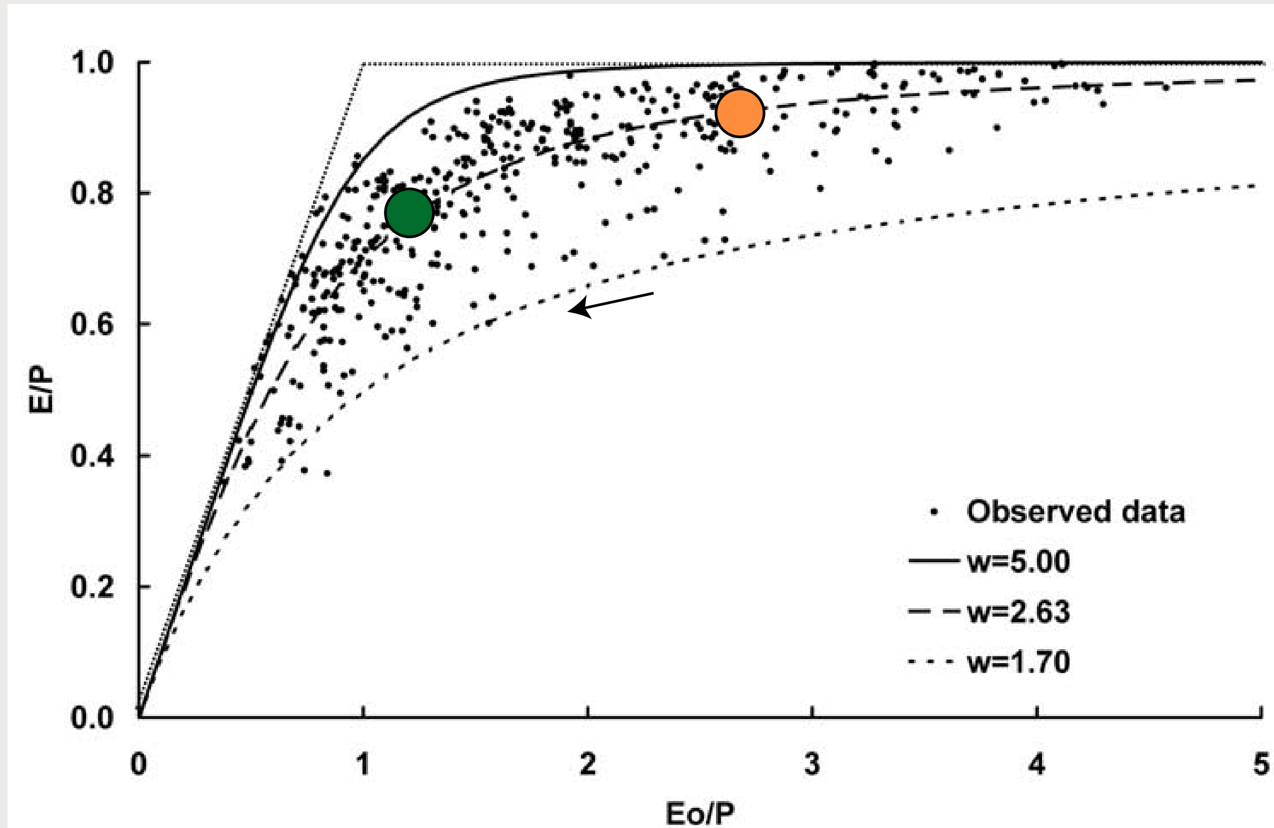
Michael Budyko, a Russian climatologist, first suggested that all catchments follow a similar behavior in the ET_{pot}/P, ET_{act}/P-space.

adapted from: Sivapalan, The Annual Water Balance

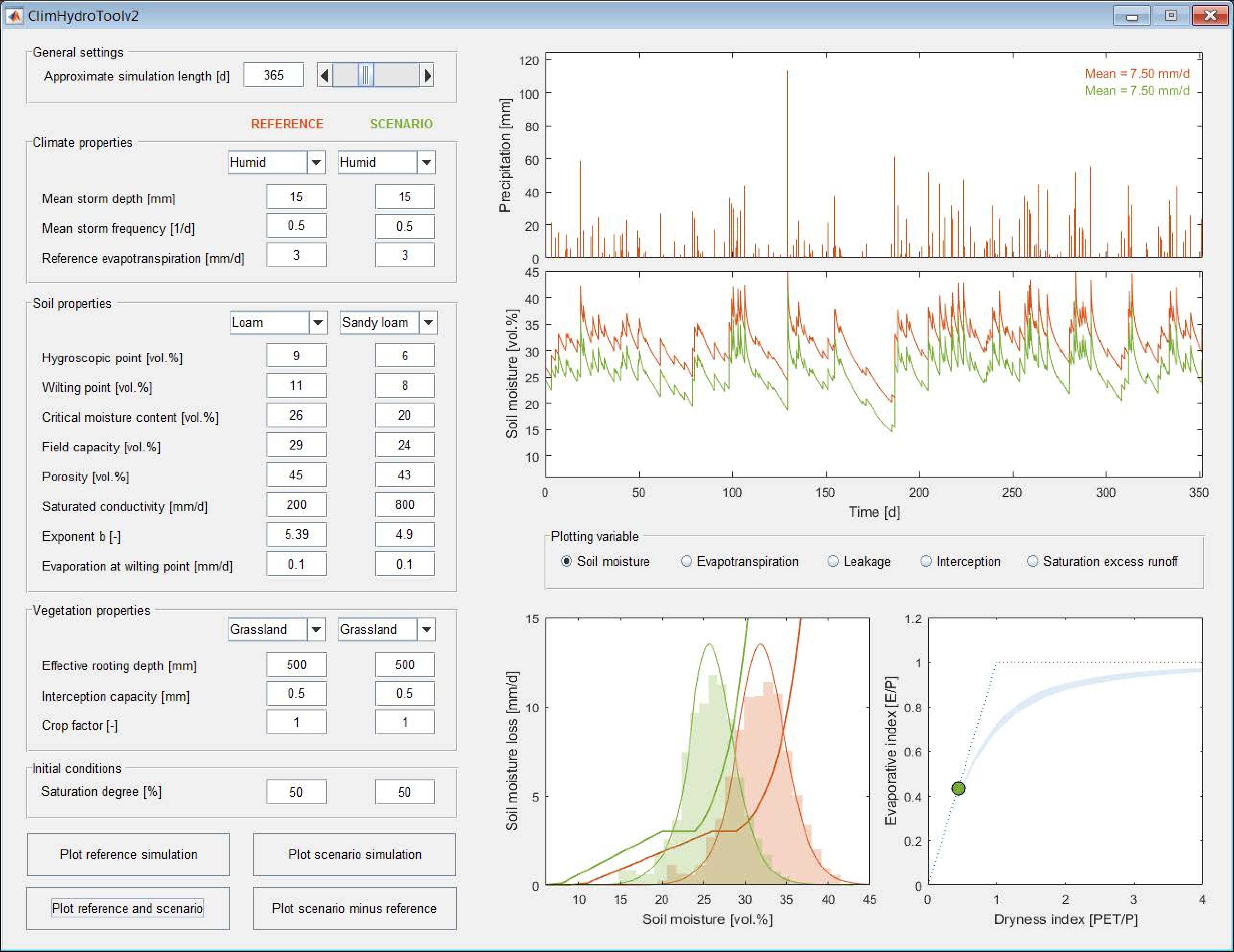
A model for the Budyko curve

$$\frac{E}{P} = 1 + \frac{E_0}{P} - \left[1 + \left(\frac{E_0}{P}\right)^{w}\right]^{1/w}$$

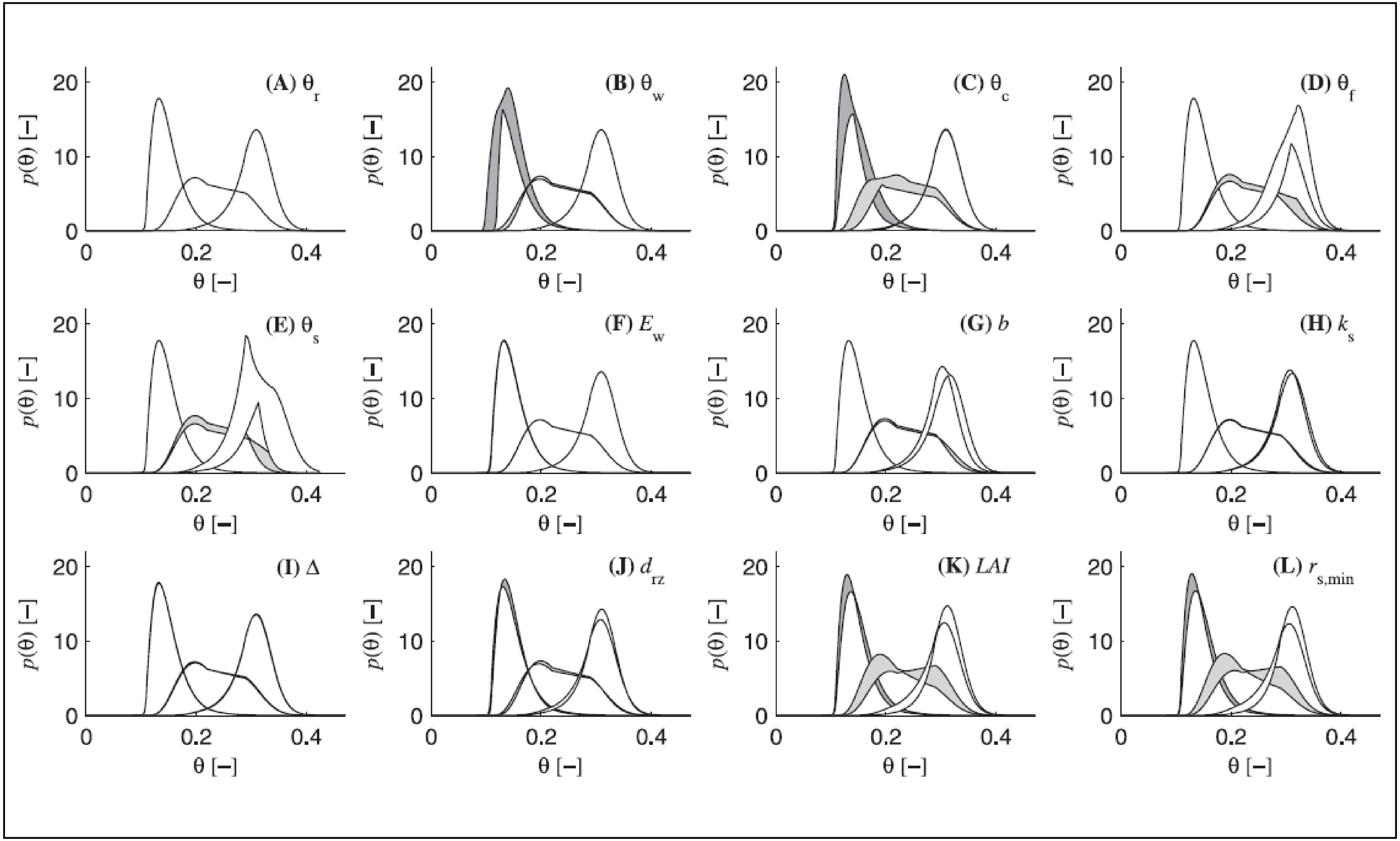
For many catchments around the world, it was found that w = 2.63. This is a first-order catchment model that can be used to predict water balance partitioning across different climate zones.



Zhang et al., 2004. A rational function approach for estimating mean annual evapotranspiration. *Water Resour. Res.* **40**


ClimHydroToolv2
General settings
Approximate simulation length [d]
365
REFERENCE
SCENARIO
Climate properties
Humid
Humid
Mean storm depth [mm]
15
15
Mean storm frequency [1/d]
0.5
0.5
Reference evapotranspiration [mm/d]
3
3
Soil properties
Loam
Sandy loam
Hygroscopic point [vol.%]
9
6
Wilting point [vol.%]
11
8
Critical moisture content [vol.%]
26
20
Field capacity [vol.%]
29
24
Porosity [vol.%]
45
43
Saturated conductivity [mm/d]
200
800
Exponent b [-]
5.39
4.9
Evaporation at wilting point [mm/d]
0.1
0.1
Vegetation properties
Grassland
Grassland
Effective rooting depth [mm]
500
500
Interception capacity [mm]
0.5
0.5
Crop factor [-]
1
1
Initial conditions
Saturation degree [%]
50
50
Plot reference simulation
Plot scenario simulation
Plot reference and scenario
Plot scenario minus reference
Mean = 7.50 mm/d
Mean = 7.50 mm/d
Precipitation [mm]
Soil moisture [vol.%]
Time [d]
Plotting variable
Soil moisture
Evapotranspiration
Leakage
Interception
Saturation excess runoff
Soil moisture loss [mm/d]
Soil moisture [vol.%]
Evaporative index [E/P]
Dryness index [PET/P]

Parameter sensitivity across climates



Teuling et al. (2009), *J. Hydrometeorol.* **10**