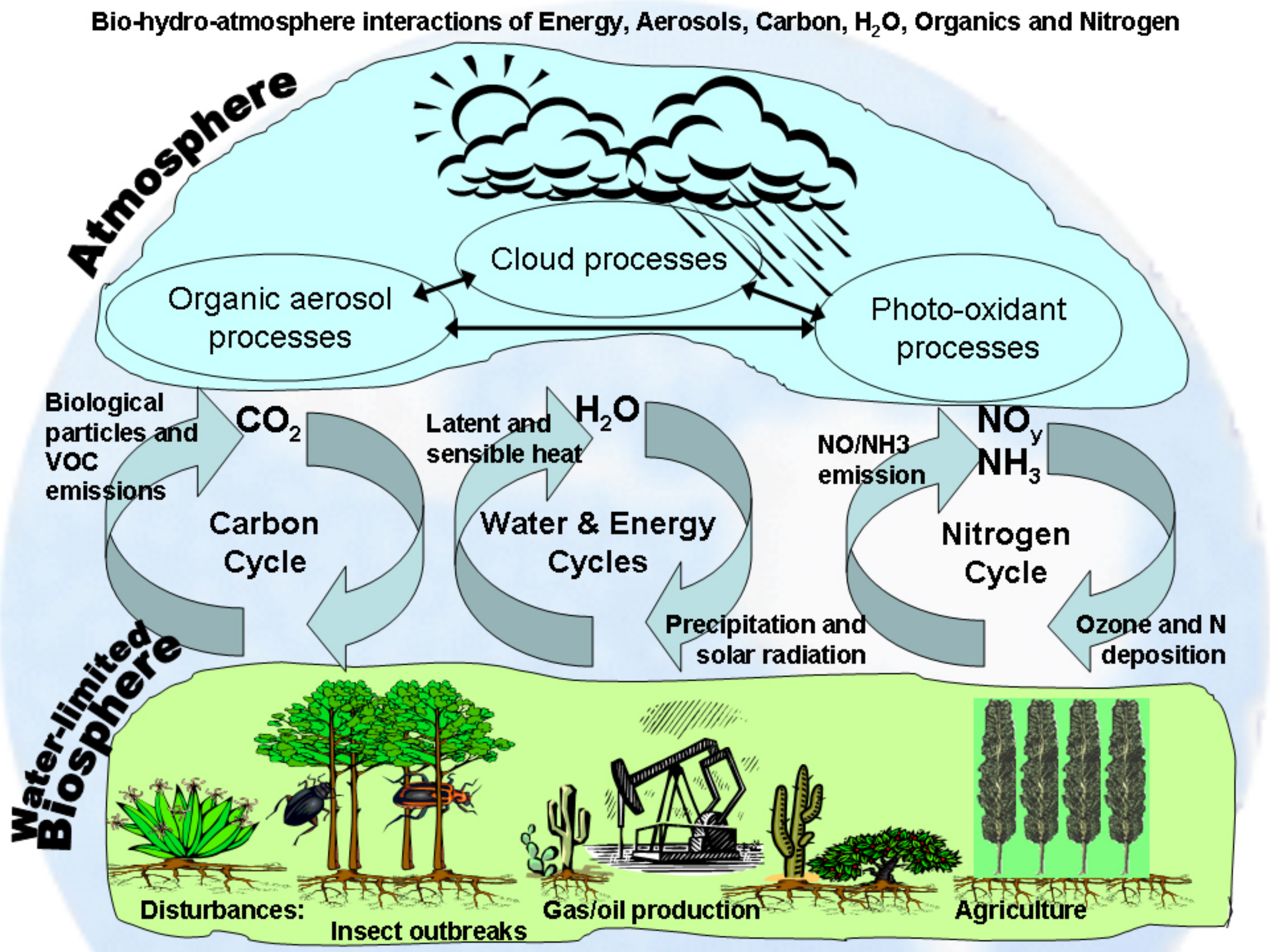
Climate in a vertical column

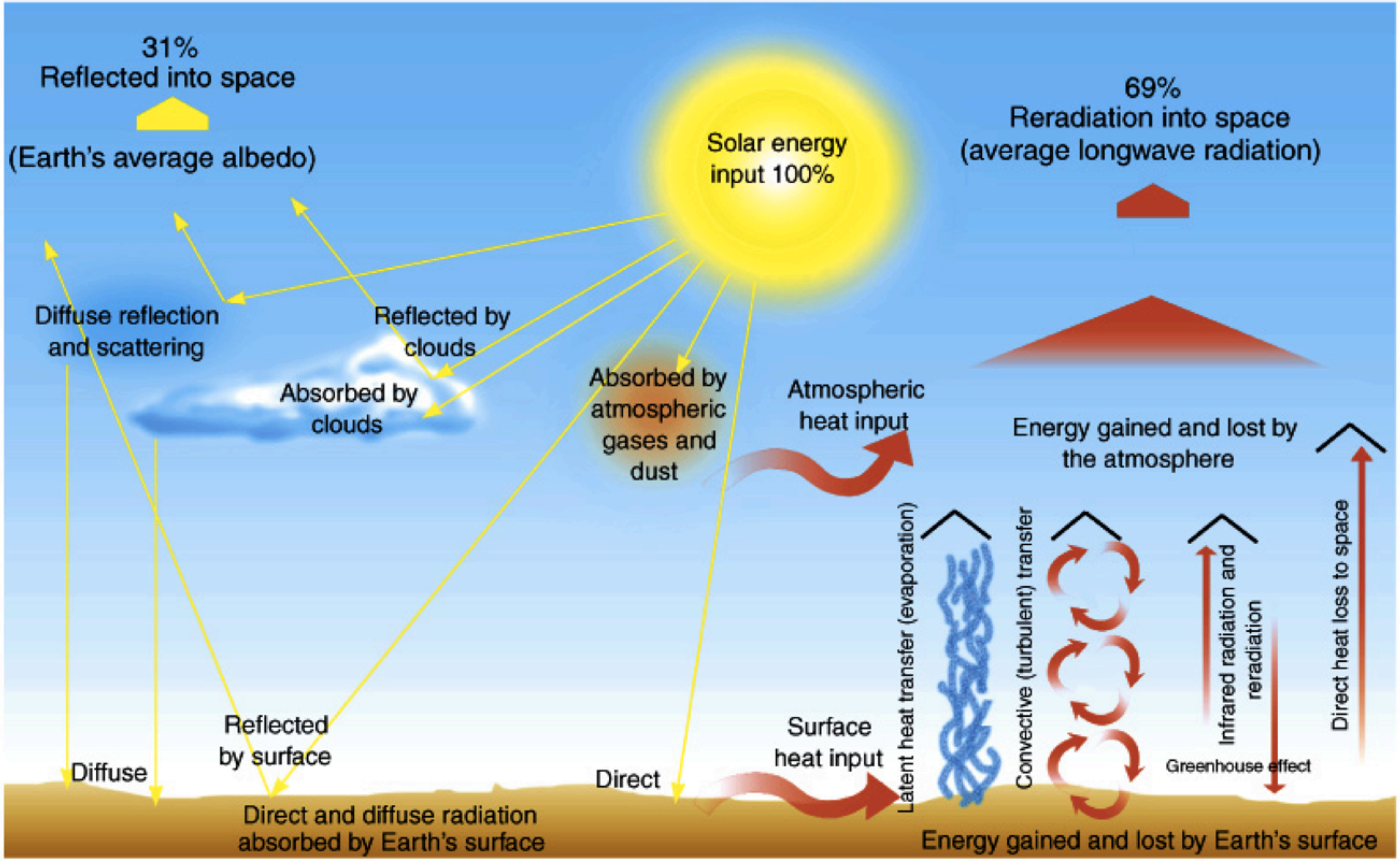
Antonello Provenzale
Istituto di Scienze dell'Atmosfera e del Clima
Consiglio Nazionale delle Ricerche

Budapest, January 2013

The interaction between soil,vegetation and atmosphere is a crucial component of the climate system



NCAR



Earth-atmosphere energy budget

Vertical energy exchanges

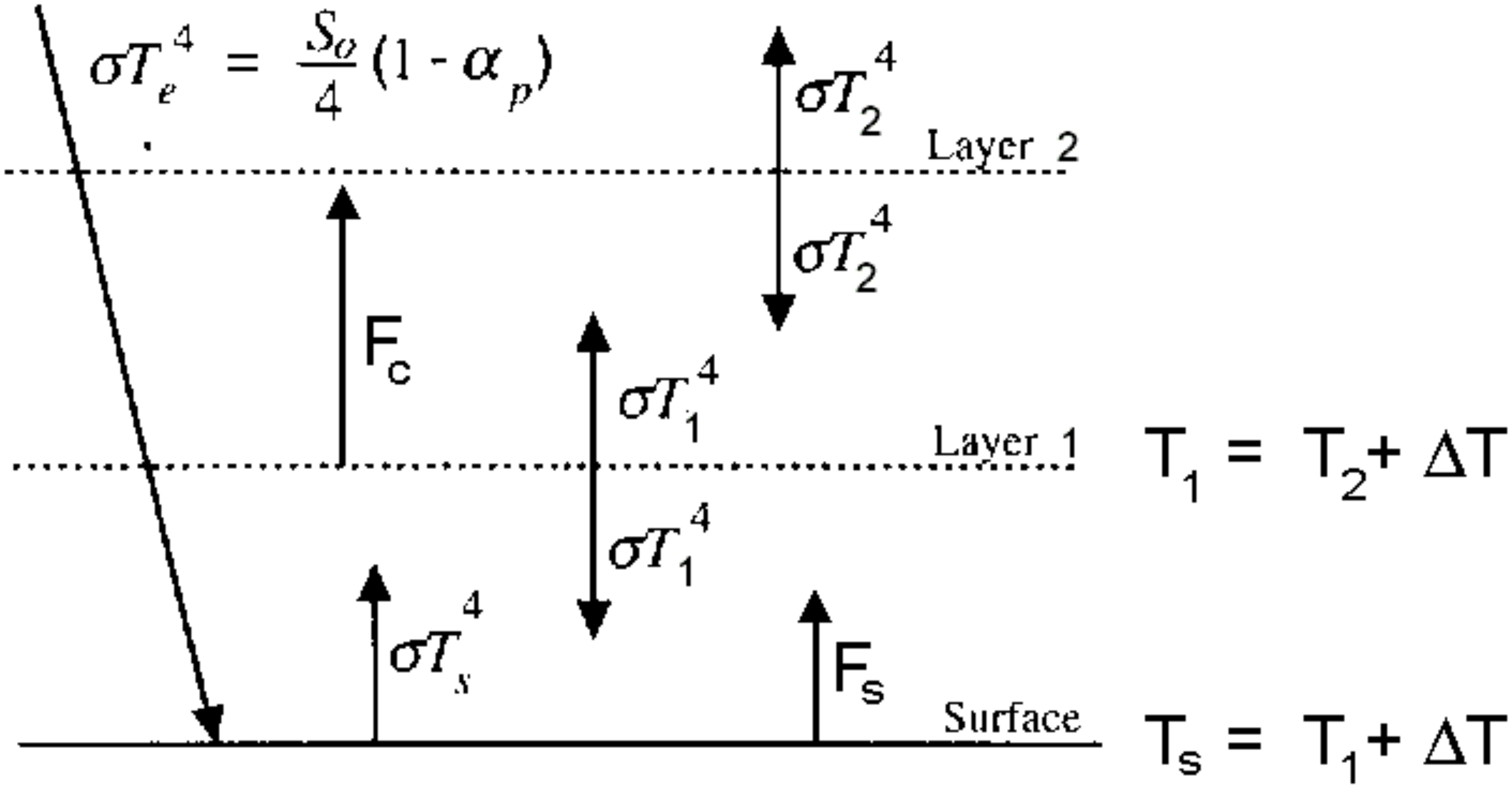
$$F_s + \sigma T_s^4 = \frac{S_0}{4}\left(1-\alpha\right) + \sigma T_1^4$$

$$F_c + 2\sigma T_1^4 = \sigma T_s^4 + \sigma T_2^4 + F_s$$

$$2\sigma T_2^4 = \sigma T_1^4 + F_c$$

$$T_1 = T_2 + \Delta T$$

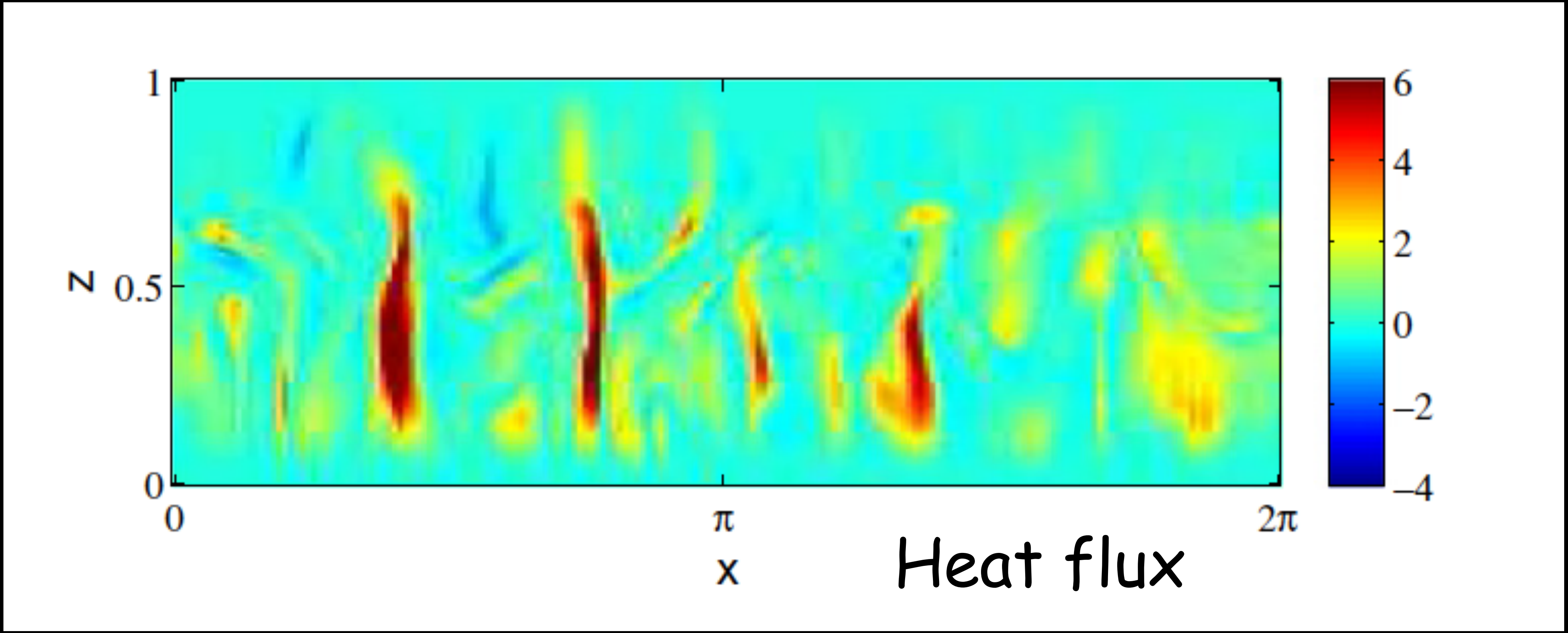
$$T_s = T_1 + \Delta T = T_2 + \Delta T$$



http://wind.mit.edu/~emanuel/geosys/node3.html

Radiative-convective column models

Of course, the situation is much more complicated



Heat flux

A conceptual model of under-saturated atmospheric convection

Berlengiero et al. 2012

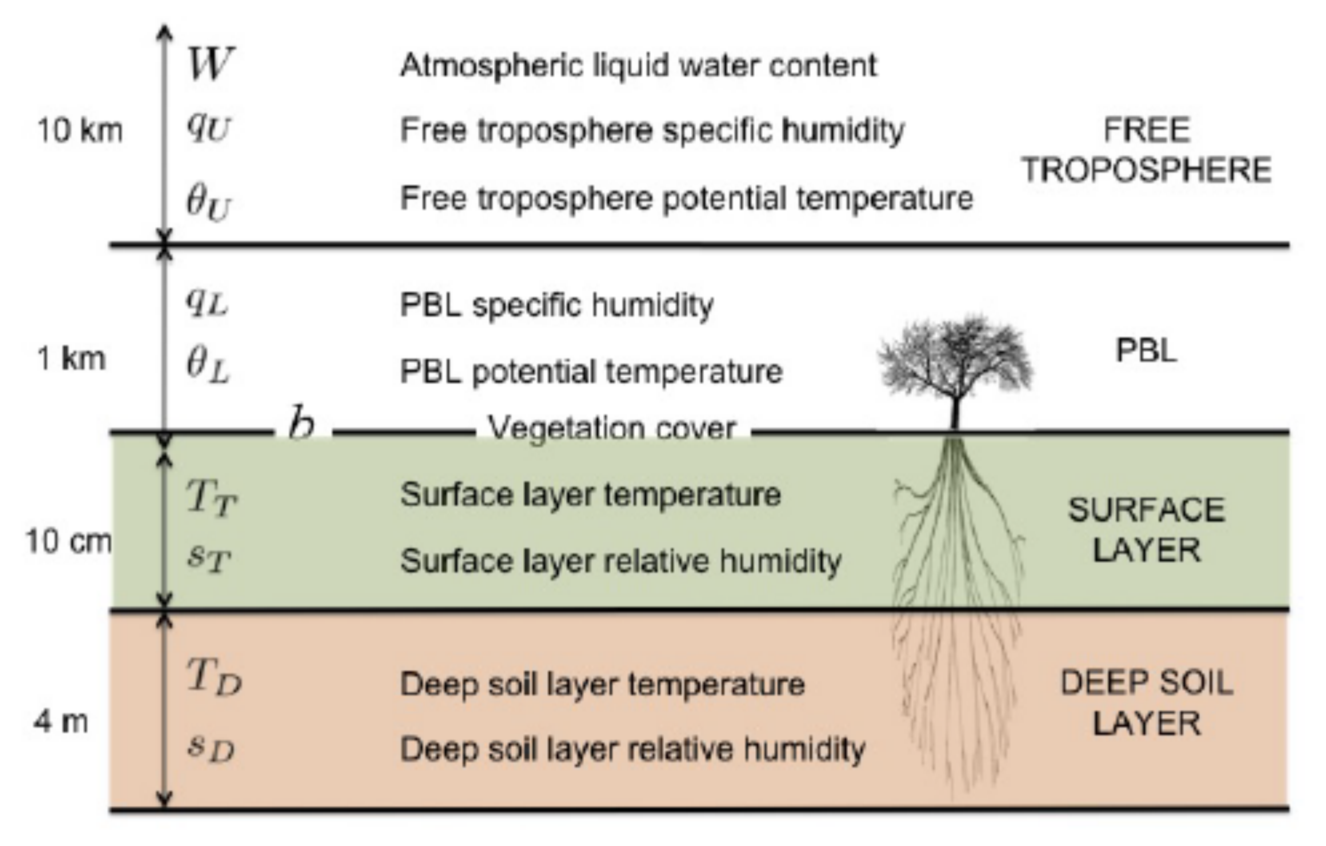
Water phase transitions further complicate the picture owing to latent heat transport:

Evaporation from the surface (surface cooling)

Upward (convective) transport

Condensation and latent heat release

Simplified models of soil-vegetation-atmosphere interactions and of vertical energy/mass transfers



1. Summer heat waves at continental midlatitudes, local water recycling and the role of vegetation

(with M. Baudena, F. D’Andrea)

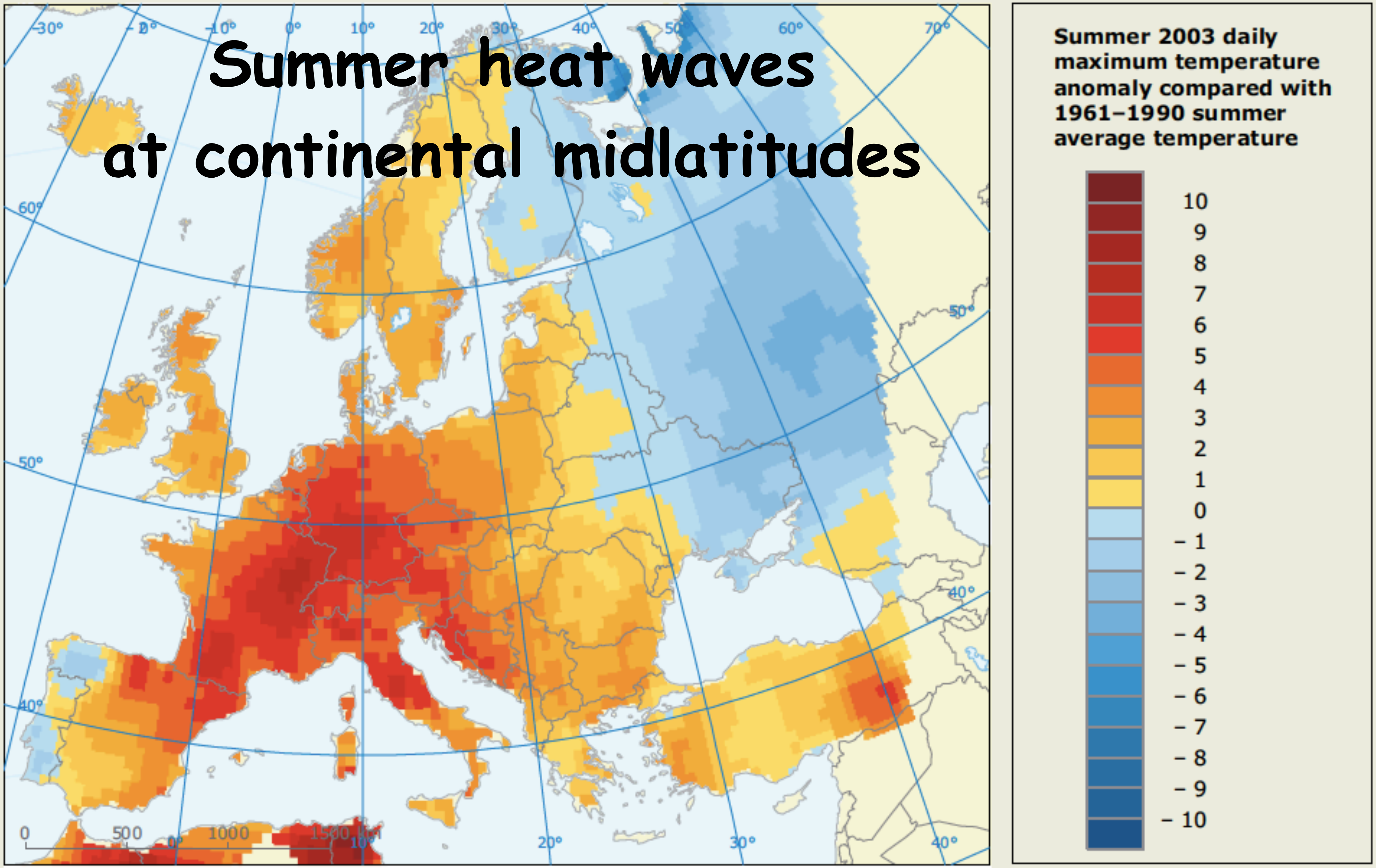
2. Planet Dune

(with F. Cresto Aleina, M. Baudena, F. D’Andrea)

3. Some thoughts on local feedbacks and upscaling

Appendix:
Vegetation patterns and moisture fluxes

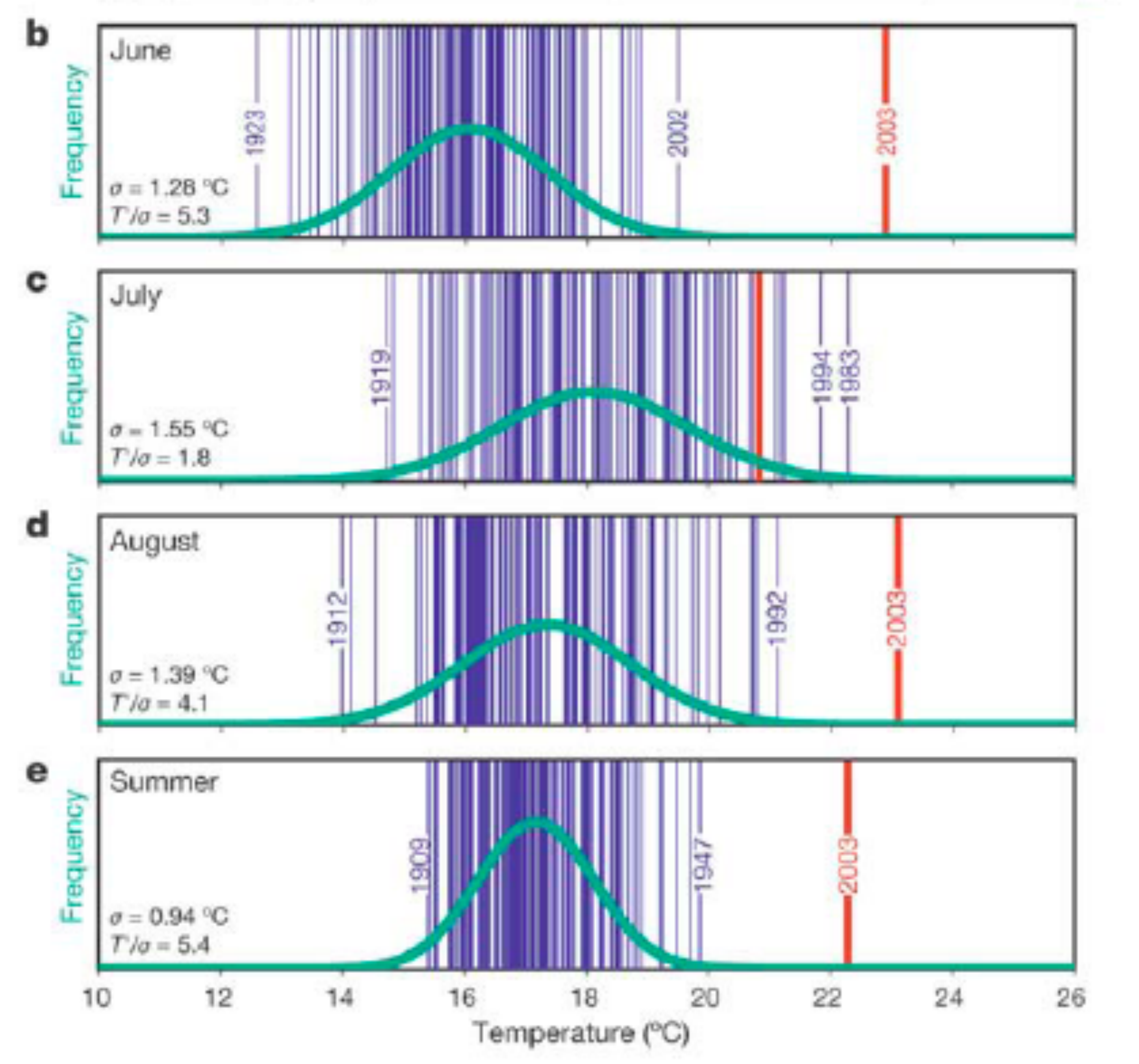
Map 5.8 Summer 2003 (June–August) daily maximum temperature anomaly



Summer heat waves at continental midlatitudes

Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

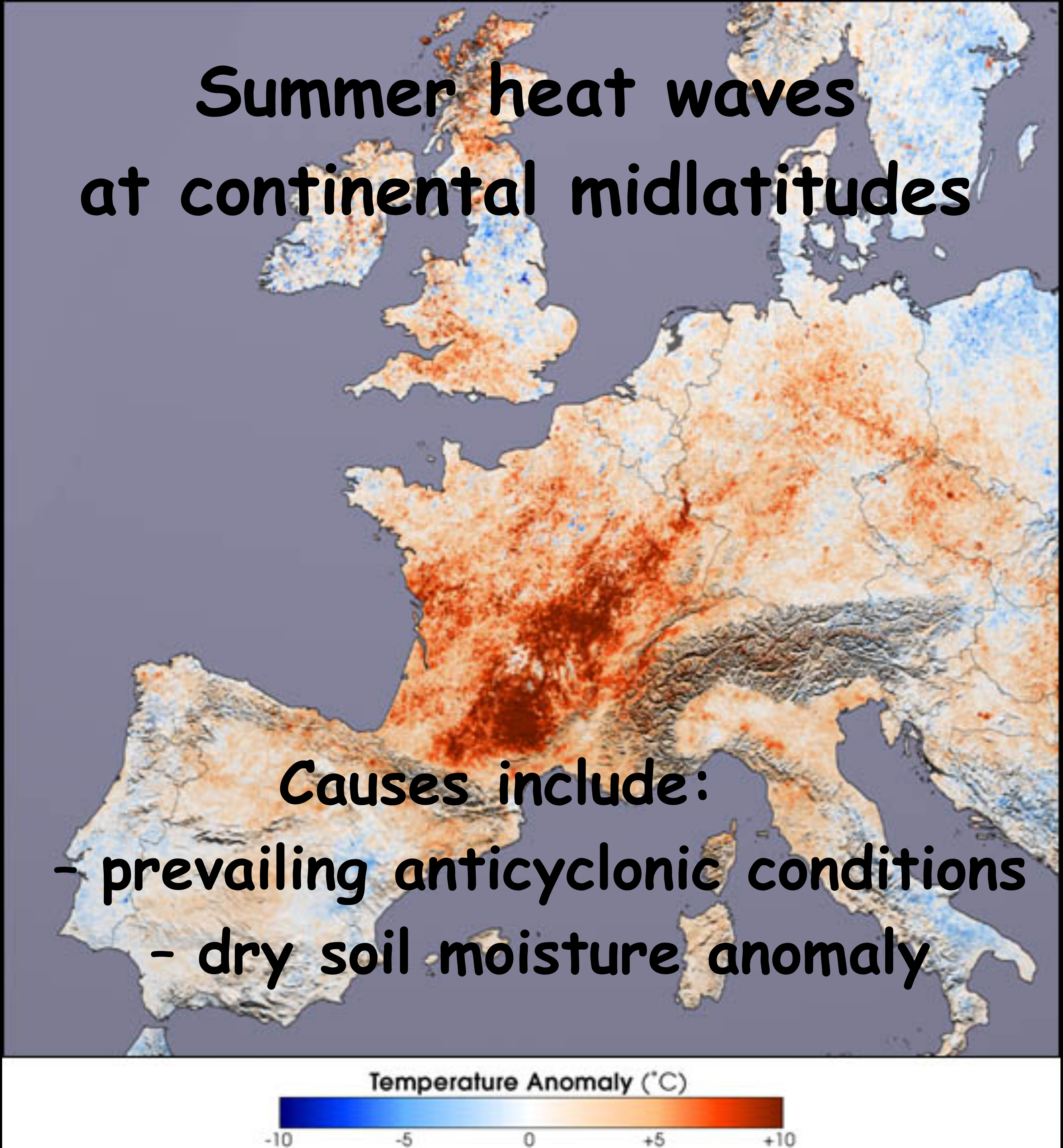
Summer 2003 : +3-5 °C above average, significant damage (mortality increase, agricultural and economic impacts)



Beniston 2004,
D'Andrea 2006,
Ferranti and Viterbo 2006,
Fisher 2007,
Meehl and Tebaldi 2004,
Schär et al. 1999,
Schär *et al.* 2004,
Stott 2004,
Vautard 2007,
Vidale 2007

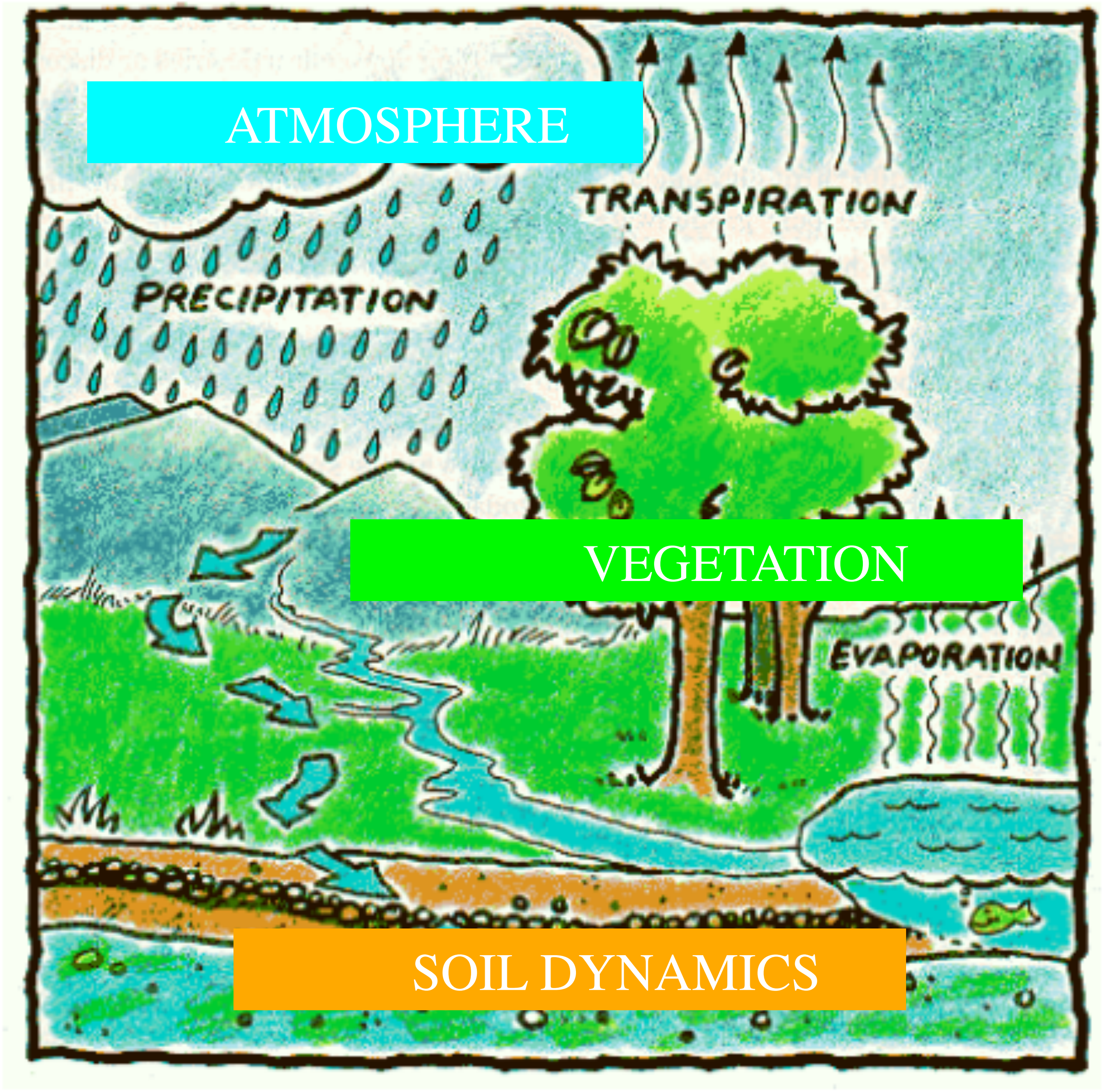
(Schär et al 2004, *Nature*)

Summer heat waves at continental midlatitudes



Causes include:

- prevailing anticyclonic conditions
- dry soil moisture anomaly

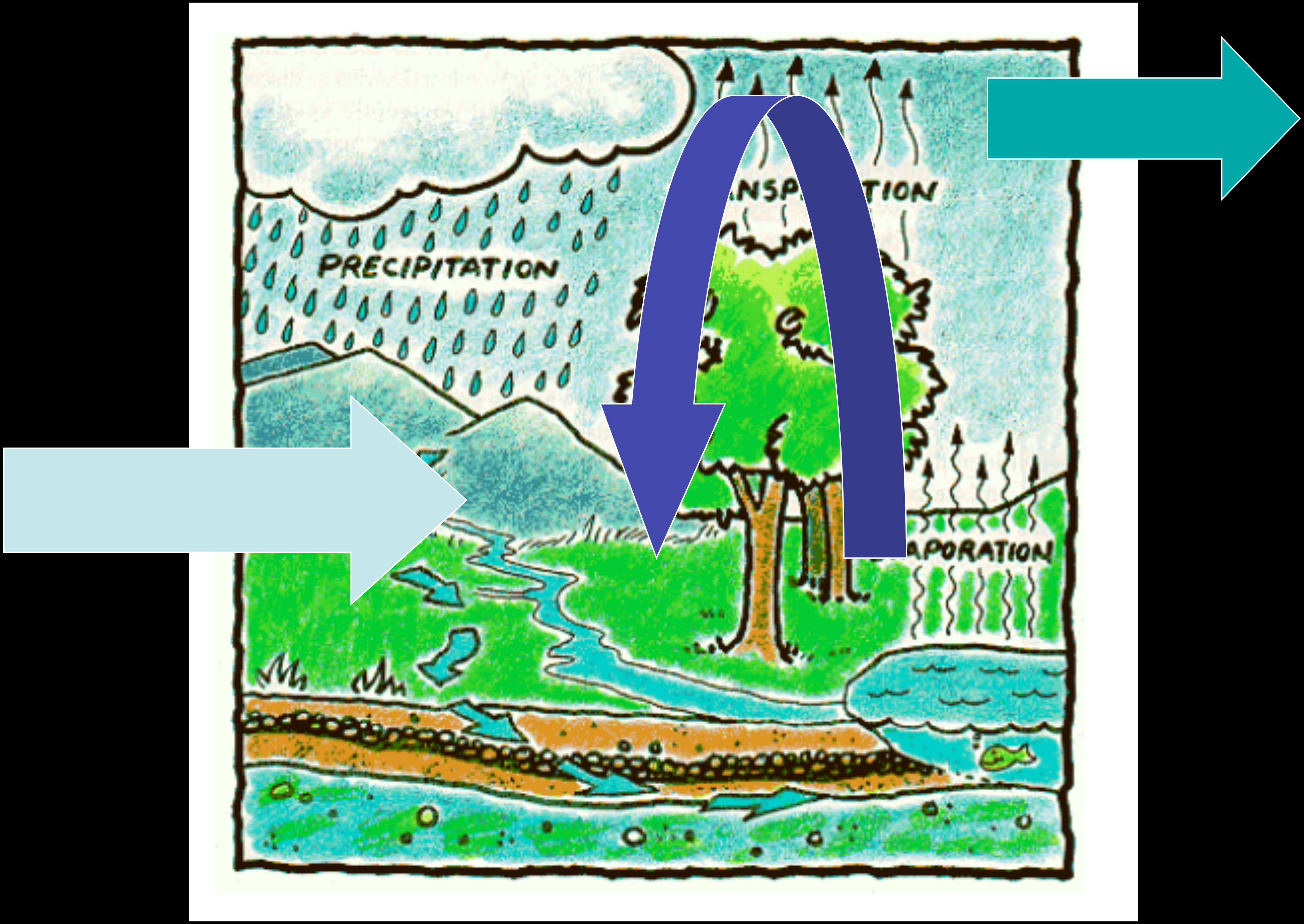


ATMOSPHERE

VEGETATION

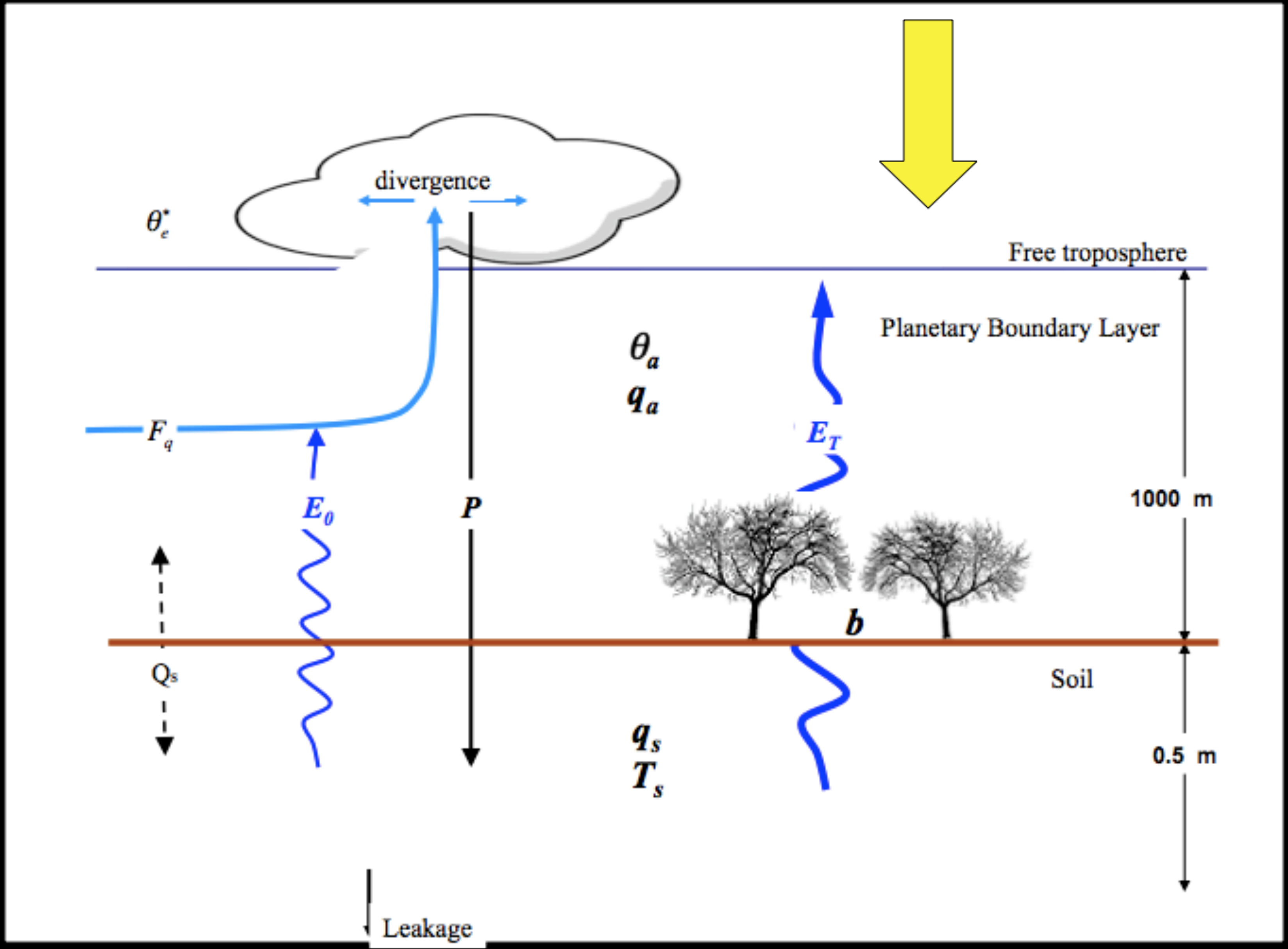
SOIL DYNAMICS

Hydrologic cycle on land



Closing the cycle: Long-range transport vs “local” recycling

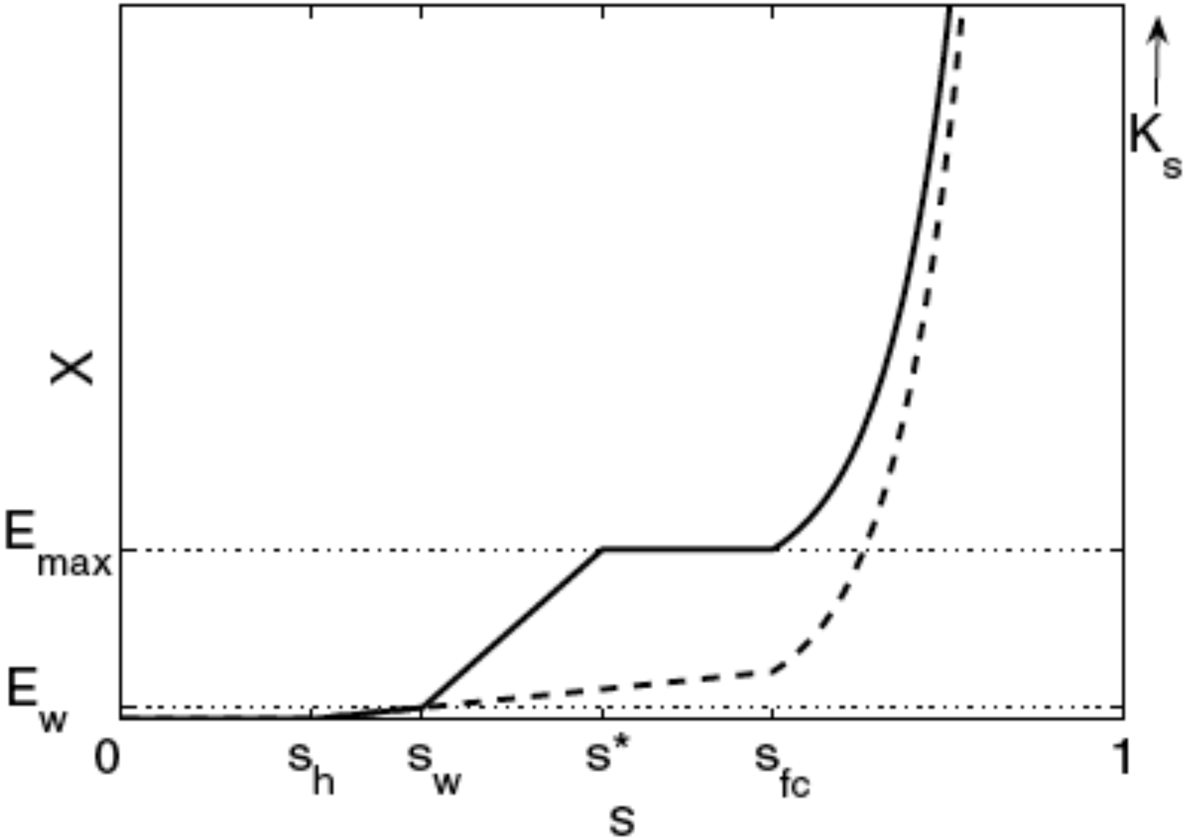
A simple box-model for the soil-vegetation-atmosphere interaction



D'Andrea, AP, Vautard, De Noblet-Ducoudrè, GRL, 2006

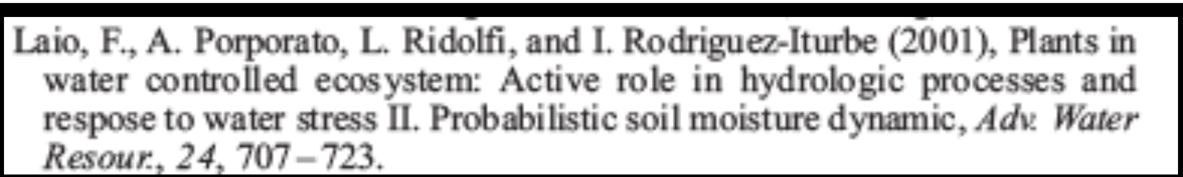
Baudena, D'Andrea, AP, WRR, 2008

$$
\begin{aligned}
\rho c_p h_a \frac{\mathrm{d}\theta_a}{\mathrm{d}t} &= \epsilon_a \epsilon_s \sigma T_s^4 + Q_s - \rho c_p h_a \frac{\mathrm{d}\Delta\tilde{\theta}_a}{\mathrm{d}t} + \frac{1}{\tau_a}(\theta_a^* - \theta_a) \\
\rho h_a \frac{\mathrm{d}q_a}{\mathrm{d}t} &= E - \rho h_a \frac{\mathrm{d}\Delta\tilde{q}_a}{\mathrm{d}t} + F_q \\
\rho_s c_{ps} h_s \frac{\mathrm{d}T_s}{\mathrm{d}t} &= (1-\alpha)F_{rad} - Q_s - \epsilon_s \sigma T_s^4 - L_e E \\
w_0 h_s \frac{\mathrm{d}q_s}{\mathrm{d}t} &= I - E - L \\
\frac{\mathrm{d}b}{\mathrm{d}t} &= gb(1-b) - \mu b
\end{aligned}
$$



Vegetation response to rainfall intermittency in drylands: Results from a simple ecohydrological box model

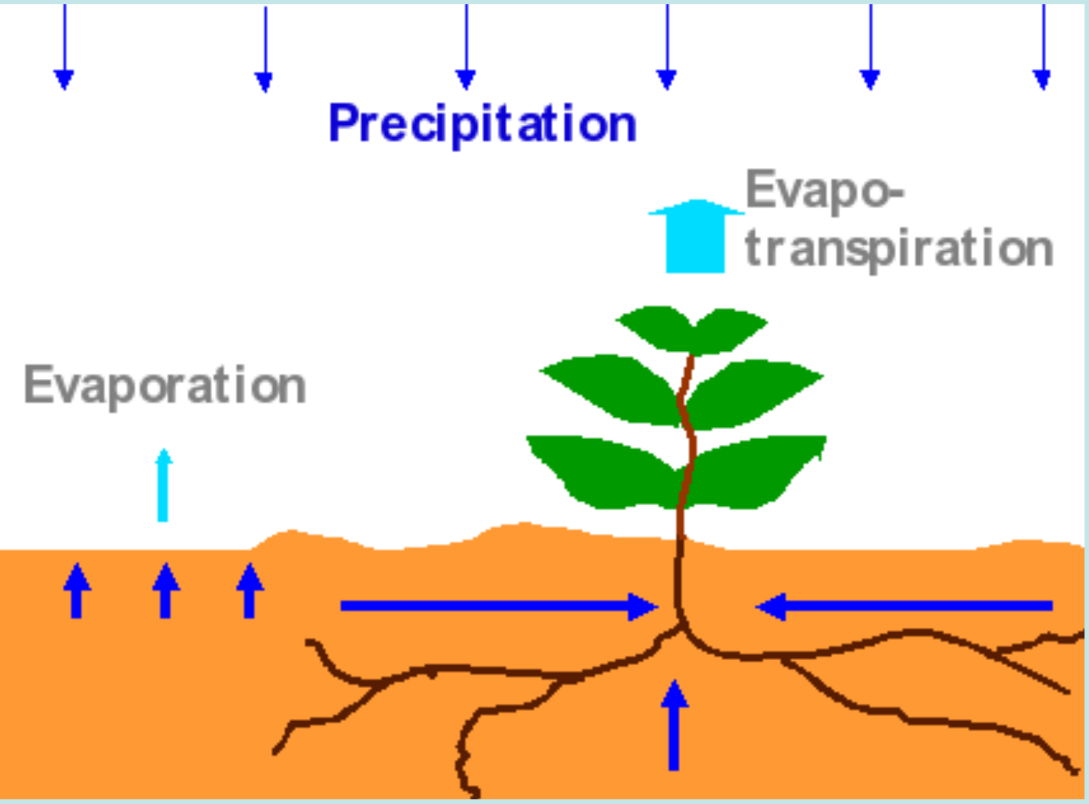
M. Baudena [a,*], G. Boni [a], L. Ferraris [a], J. von Hardenberg [b], A. Provenzale [b]

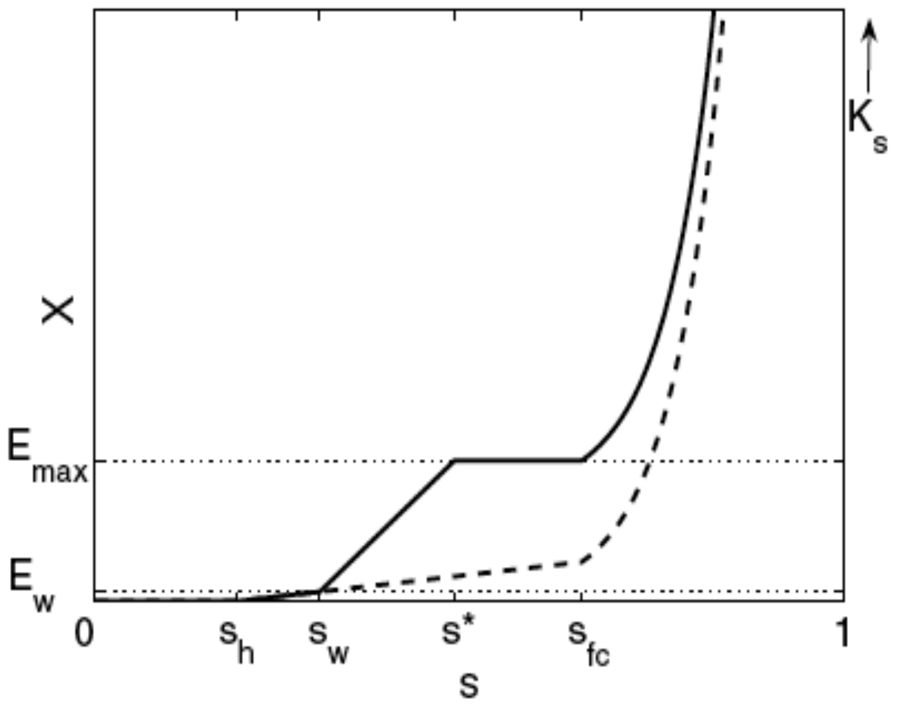


Laio, F., A. Porporato, L. Ridolfi, and I. Rodriguez-Iturbe (2001), Plants in water controlled ecosystem: Active role in hydrologic processes and respose to water stress II. Probabilistic soil moisture dynamic, *Adv. Water Resour.*, *24*, 707–723.

Rodriguez-Iturbe, I., and A. Porporato (2004), *Echohydrology of Water Controlled Ecosystems*, Cambridge Univ. Press, New York.

Evapotranspiration





Albedo

$$\alpha = b\alpha_b + (1-b)\alpha_0$$

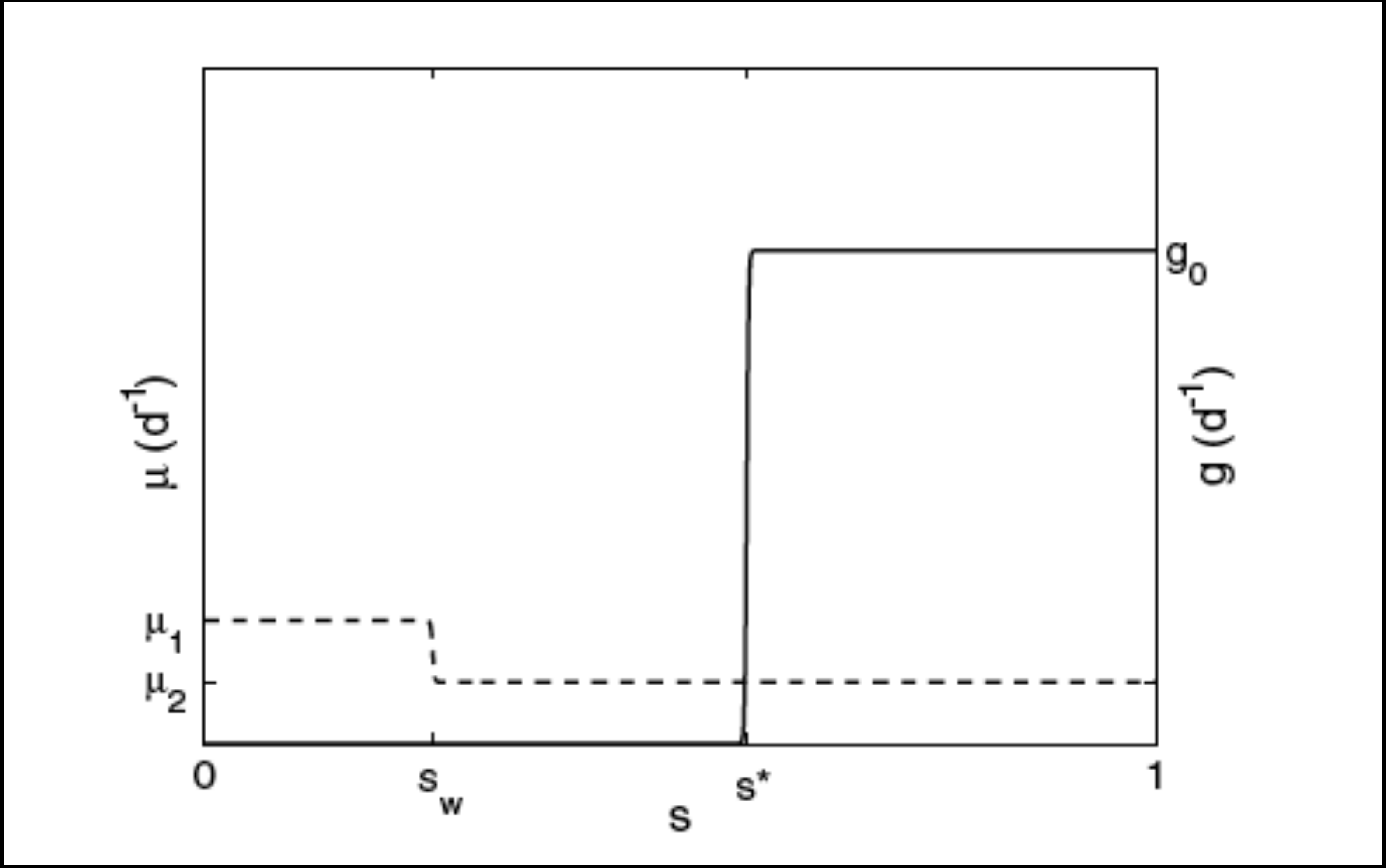
$$\alpha_0 = 0.35$$
$$\alpha_b = 0.14$$

As in Charney [1975]

Vegetation dynamics

Levins, *Bull. Entomol. Soc. Am.* 1969; Tilman, *Ecology* 1994

$$\frac{\mathrm{d}b}{\mathrm{d}t} = gb(1-b) - \mu b.$$

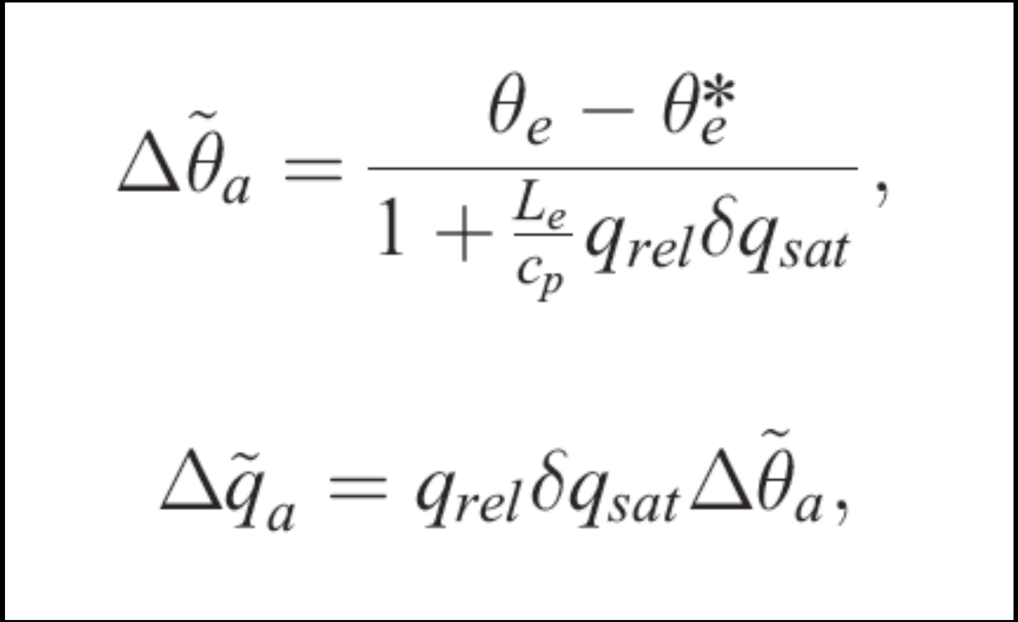


Baudena, AP, HESS 2008

Convection parameterization:

If $\theta_e = \theta_a \exp \frac{L_e q_a}{c_p \theta_a} > \theta_e^*$ convection occurs

We assume relative humidity is conserved and that convection is instantaneous

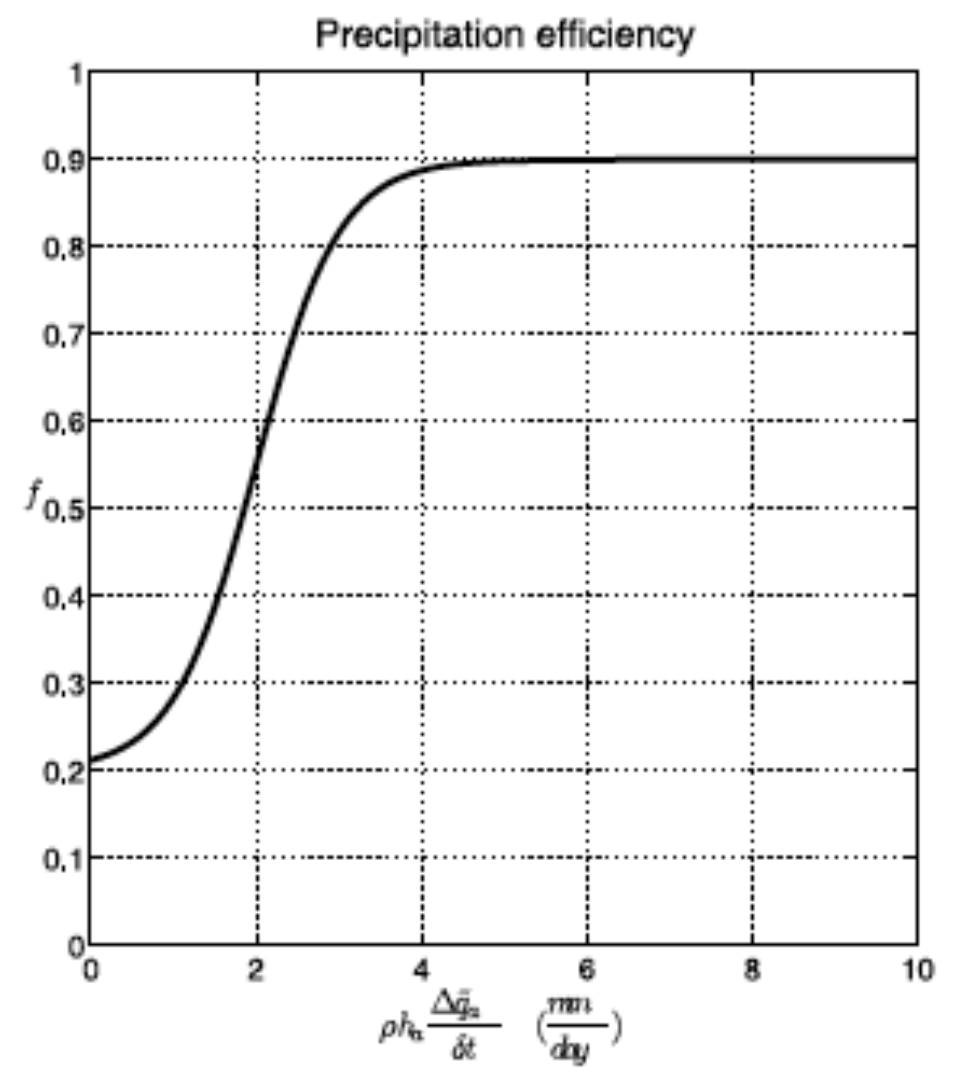


$$\Delta\tilde{\theta}_a = \frac{\theta_e - \theta_e^*}{1 + \frac{L_e}{c_p} q_{rel} \delta q_{sat}},$$

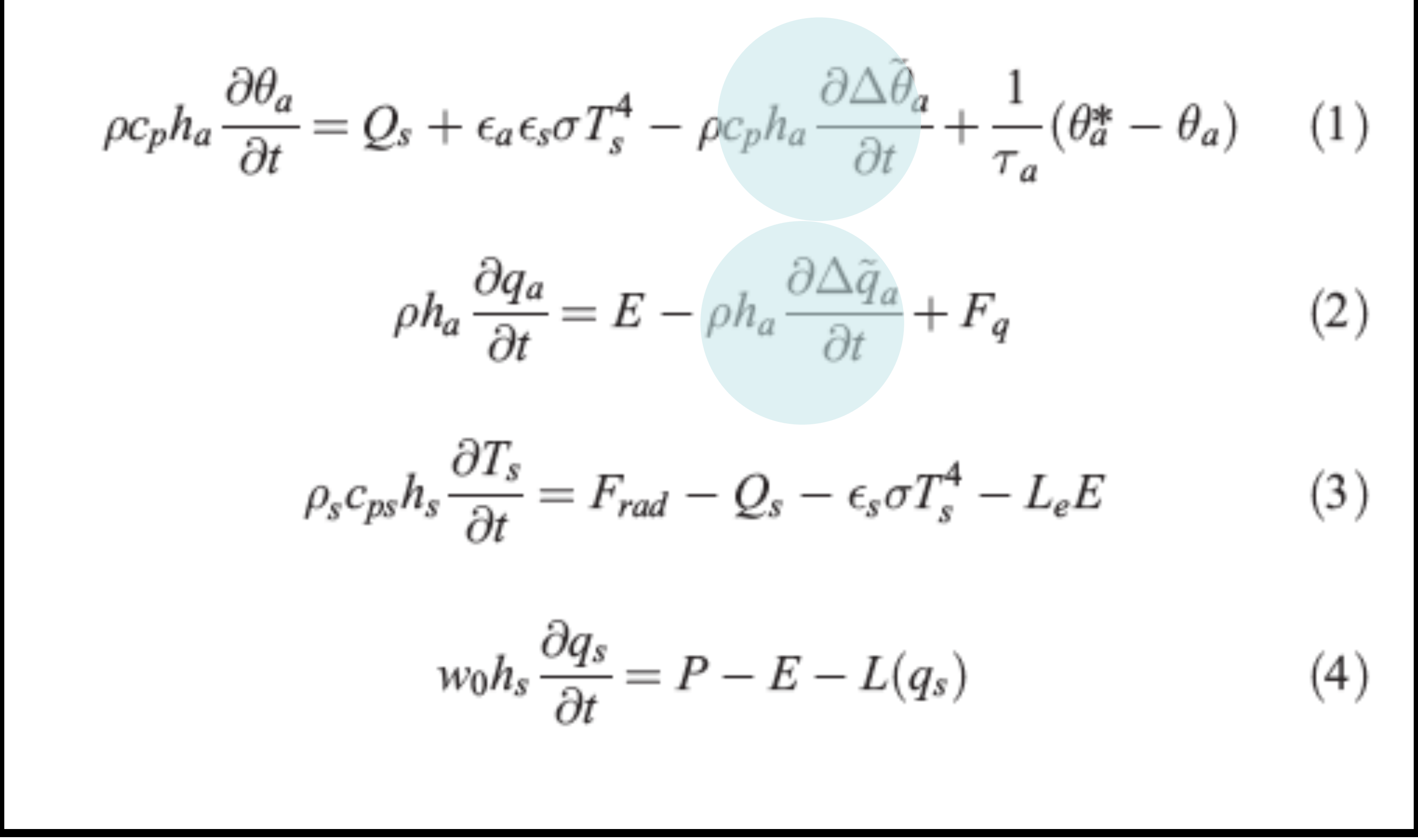
$$\Delta\tilde{q}_a = q_{rel} \delta q_{sat} \Delta\tilde{\theta}_a,$$

Precipitation efficiency (crucial feedback):

The amount of moisture transformed
into local precipitation depends on the intensity of convection



A case without vegetation dynamics (D’Andrea et al, GRL 2006)

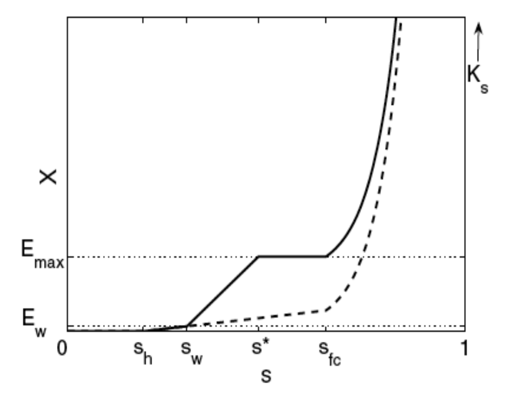


$$\rho c_p h_a \frac{\partial \theta_a}{\partial t} = Q_s + \epsilon_a \epsilon_s \sigma T_s^4 - \rho c_p h_a \frac{\partial \Delta \tilde{\theta}_a}{\partial t} + \frac{1}{\tau_a}(\theta_a^* - \theta_a) \quad (1)$$

$$\rho h_a \frac{\partial q_a}{\partial t} = E - \rho h_a \frac{\partial \Delta \tilde{q}_a}{\partial t} + F_q \quad (2)$$

$$\rho_s c_{ps} h_s \frac{\partial T_s}{\partial t} = F_{rad} - Q_s - \epsilon_s \sigma T_s^4 - L_e E \quad (3)$$

$$w_0 h_s \frac{\partial q_s}{\partial t} = P - E - L(q_s) \quad (4)$$



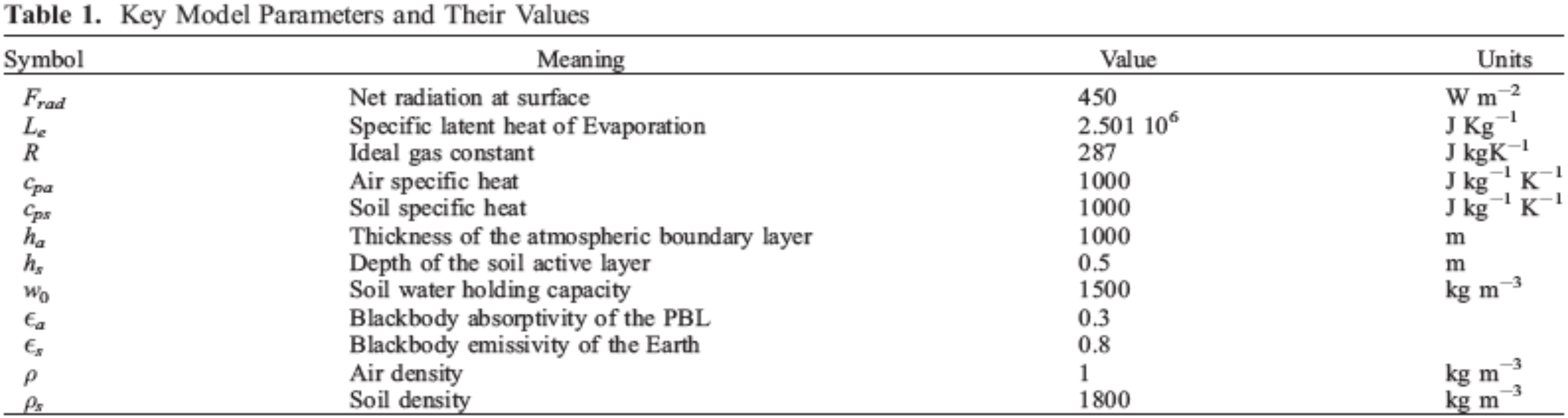


Table 1. Key Model Parameters and Their Values

Symbol	Meaning	Value	Units
F_{rad}	Net radiation at surface	450	W m^{-2}
L_e	Specific latent heat of Evaporation	2.501 10^6	J Kg^{-1}
R	Ideal gas constant	287	J kgK^{-1}
c_{pa}	Air specific heat	1000	J kg^{-1} K^{-1}
c_{ps}	Soil specific heat	1000	J kg^{-1} K^{-1}
h_a	Thickness of the atmospheric boundary layer	1000	m
h_s	Depth of the soil active layer	0.5	m
w_0	Soil water holding capacity	1500	kg m^{-3}
ϵ_a	Blackbody absorptivity of the PBL	0.3	
ϵ_s	Blackbody emissivity of the Earth	0.8	
ρ	Air density	1	kg m^{-3}
ρ_s	Soil density	1800	kg m^{-3}

Multiple equilibria of the soil-atmosphere system

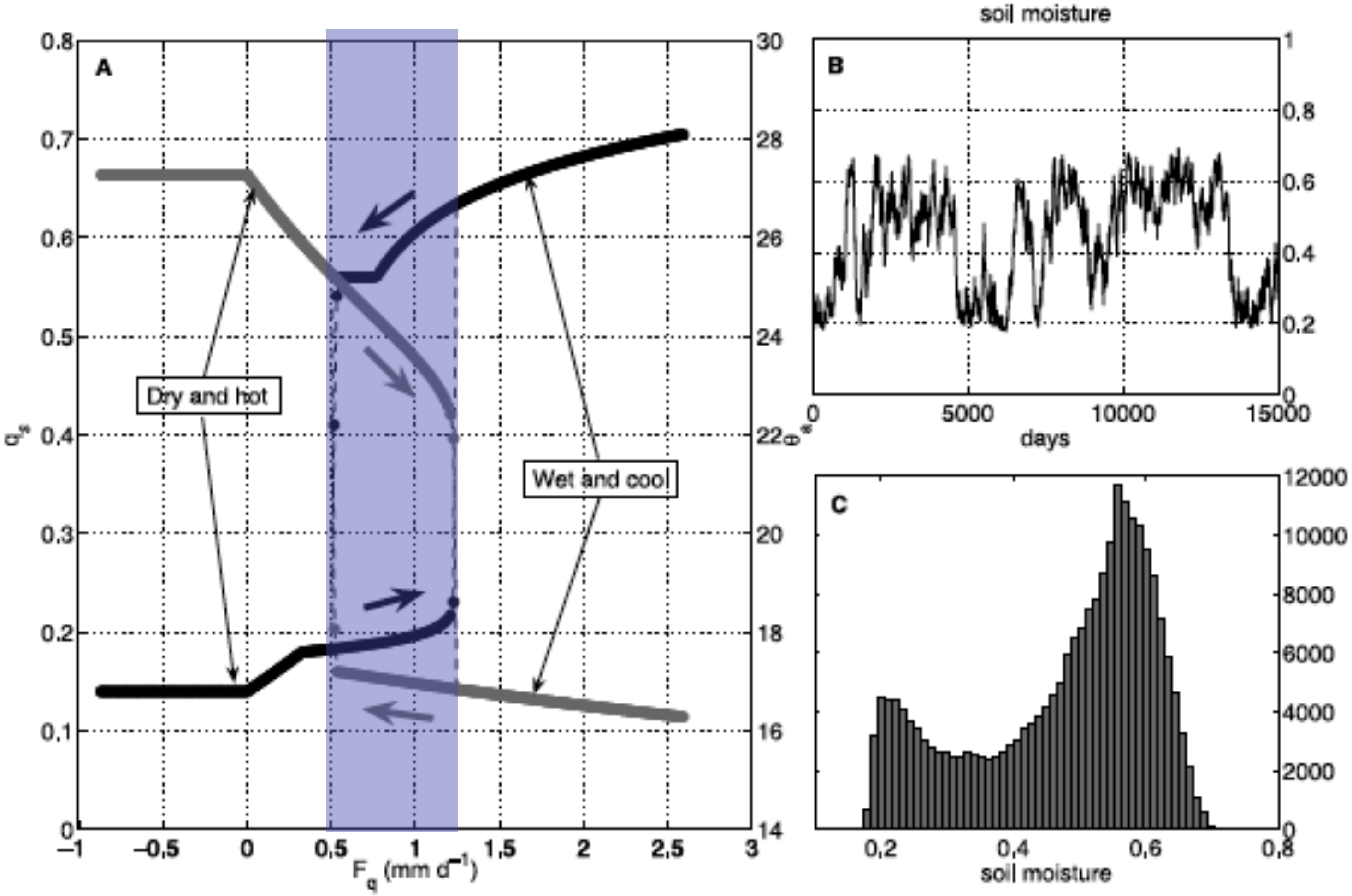
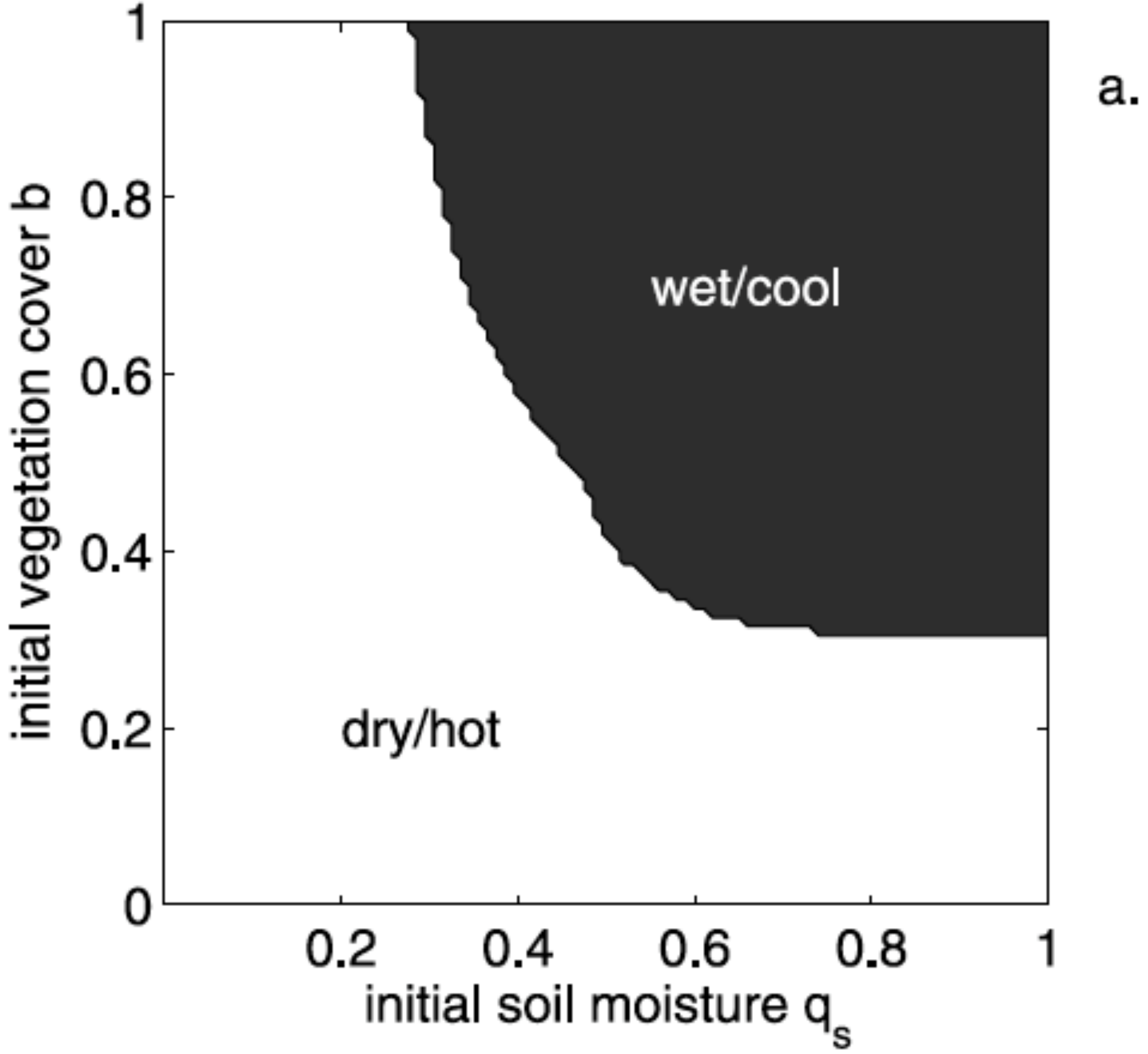
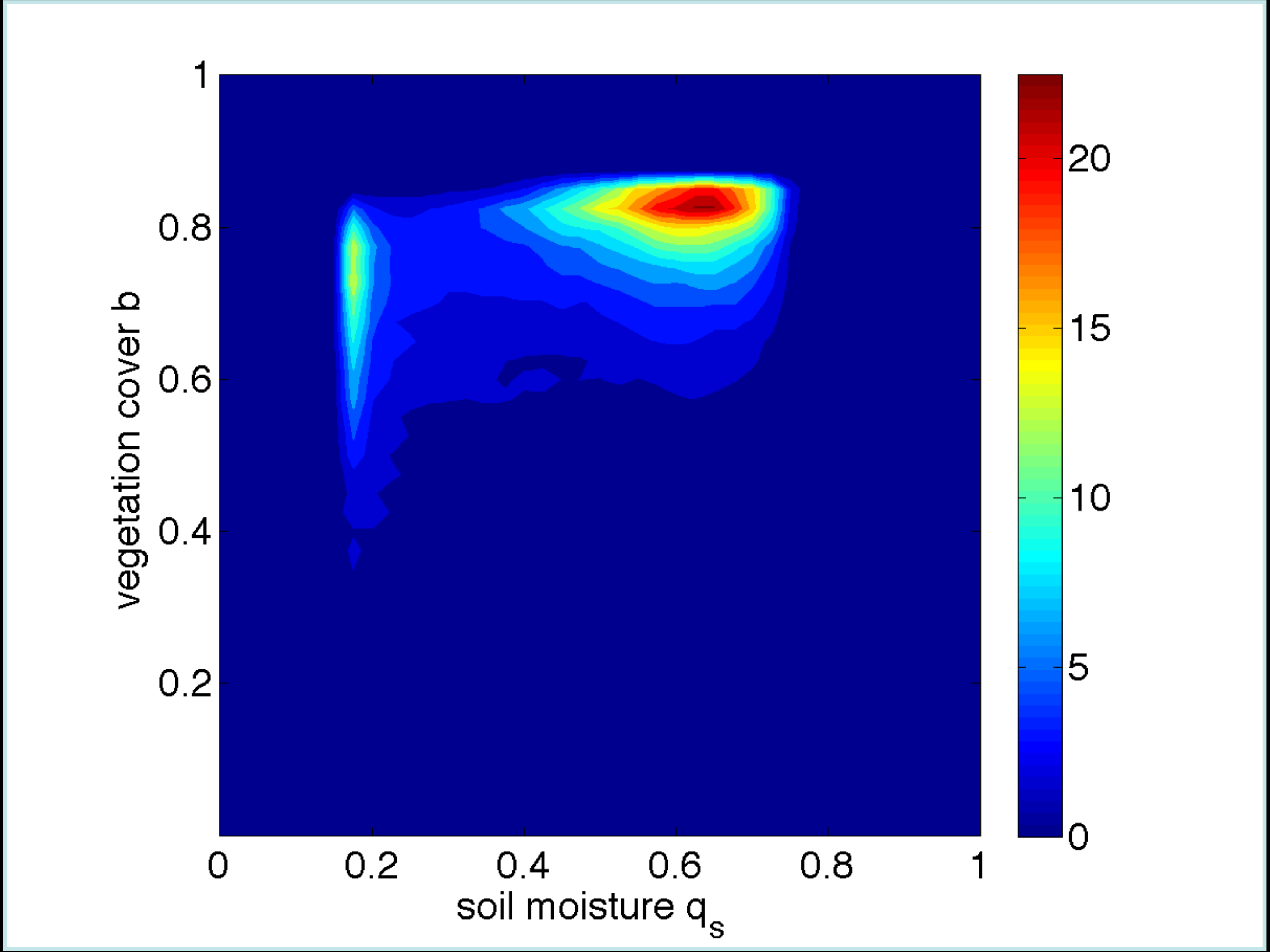


Figure 4. Sensitivity to the flux convergence F_q. (a) Values of PBL temperature (gray) and soil moisture (black) as a function of the F_q, expressed in mmd^{-1}. The branches corresponding to the dry state are attained by increasing values of F_q and the wet branches for decreasing values. Arrows mark the sense of the hysteresis cycle. (b) Time evolution of soil moisture in a 15000-day-long section of the stochastic run of the model. (c) Histogram of the values of q_s of the whole integration.

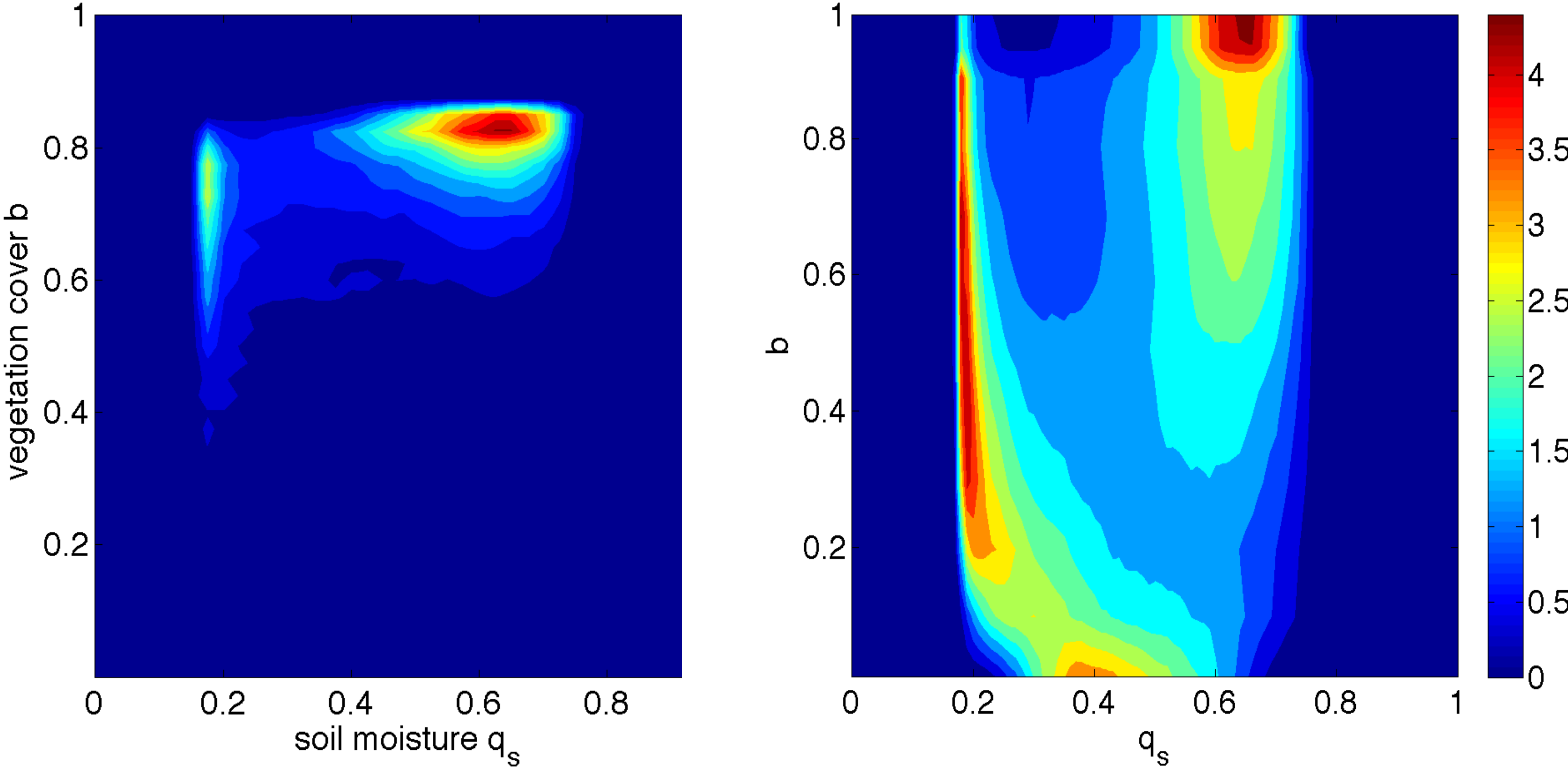


If we include vegetation

Effects of stochastic variability in F_q



Dynamical vegetation vs fixed veggies:



Hints:

Insurgence of summer droughts much dependent
on the soil-moisture conditions at the end of spring

For realistic parameter values one obtains
a bimodal distribution of soil moisture values

Vegetation cover is important:
below a minimal vegetation cover droughts are more probable

With dynamic vegetation, the “preferred” state
is moist summers. When vegetation is frozen,
summer droughts become more probable.

The crucial feedback is related to precipitation efficiency

Evapotranspiration feedback more important than albedo feedback

The climate of Dune (Planet Arrakis)

F. Cresto Aleina, M. Baudena, F. D’Andrea, AP, *Tellus B* 2013

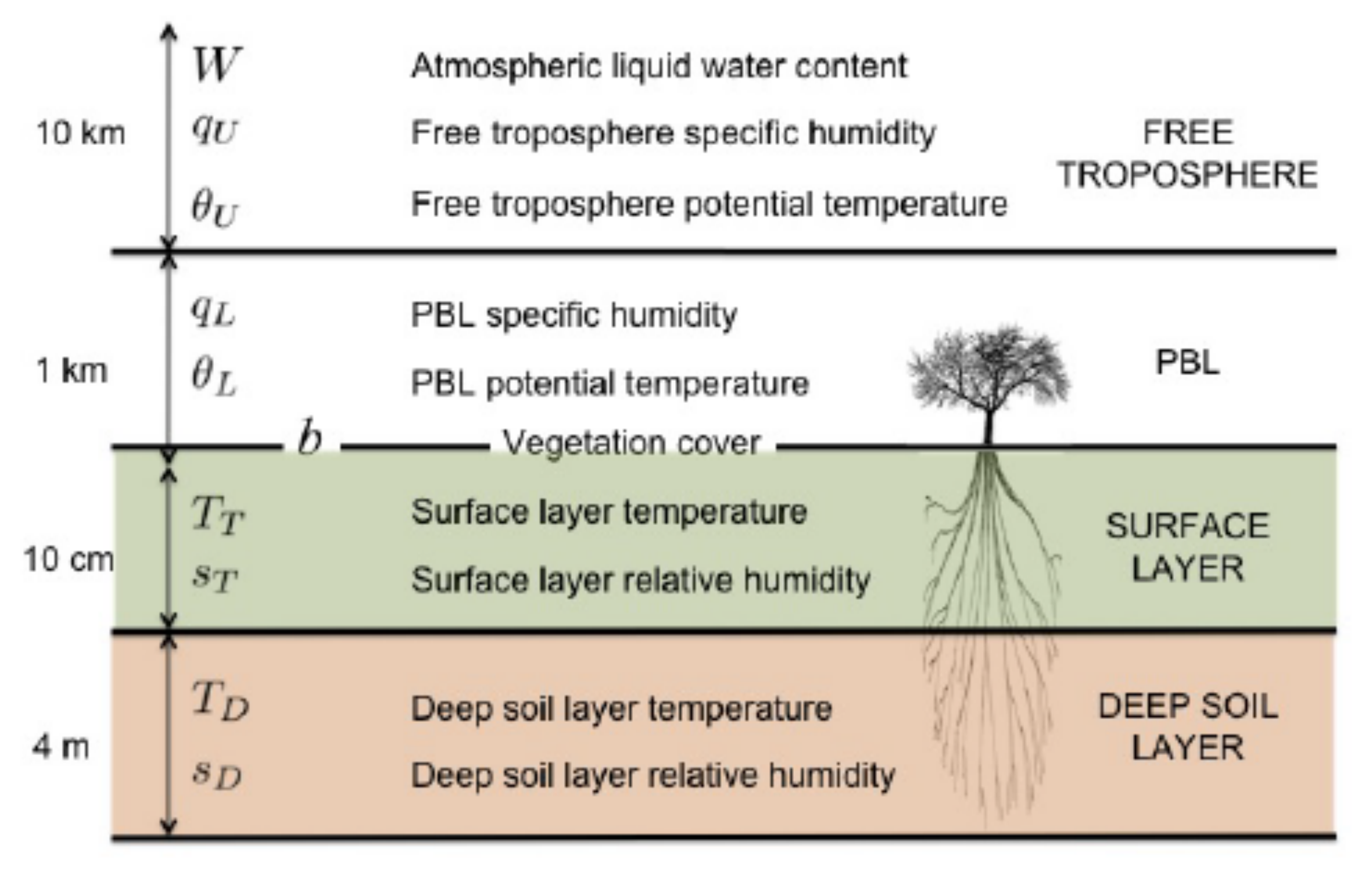
Imagine a sandy planet with no ocean; water is in the sand

No vegetation → only evaporation

Could transpiration from vegetation
generate a full hydrological cycle ?

On Earth, could transpiration from vegetated areas generate an hydrologic cycle and multiple stable states in continental areas with little moisture influx from the ocean ?

A box model for a sandy planet
(or a closed continental region)

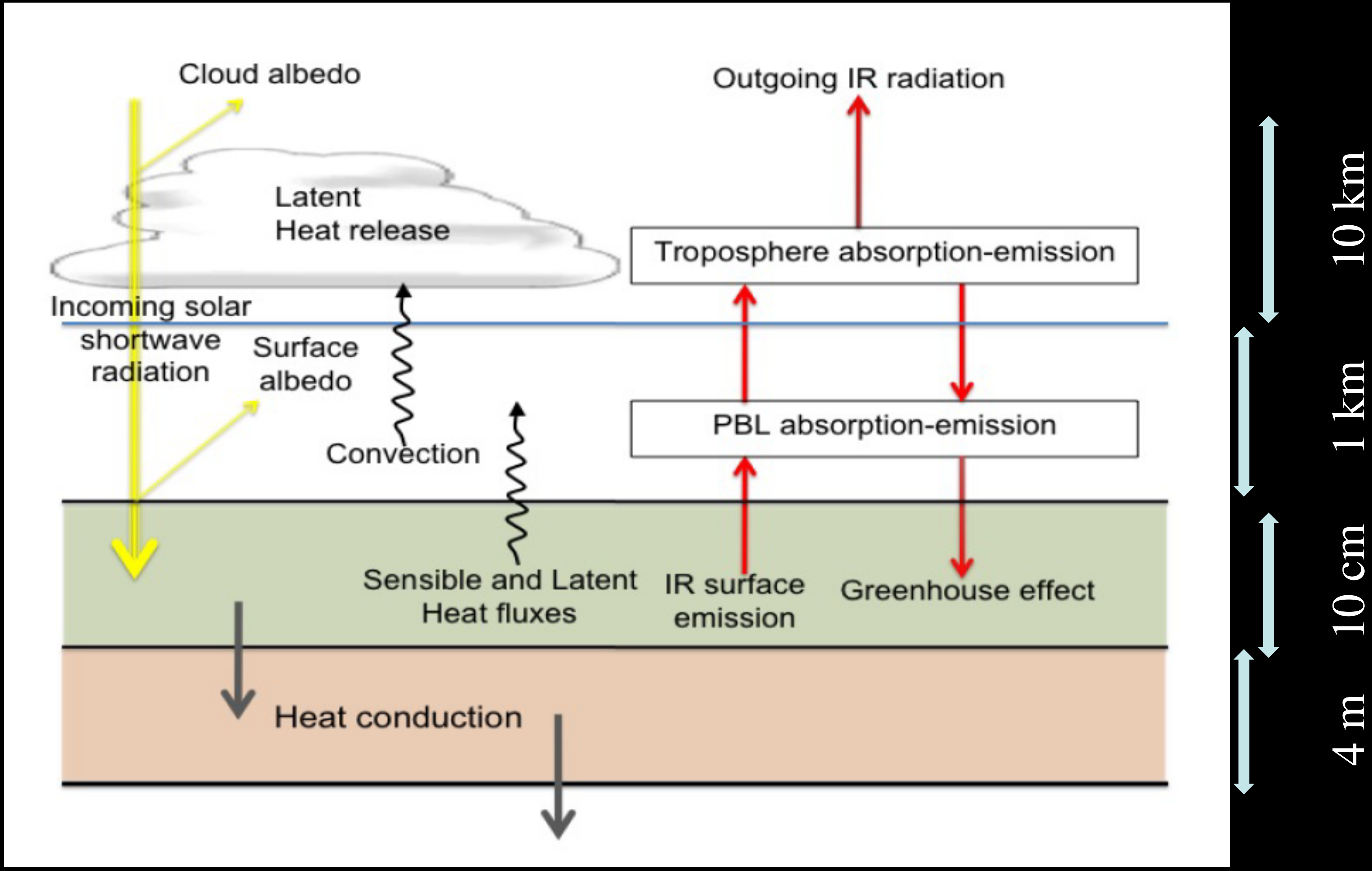


10 km	W	Atmospheric liquid water content	FREE TROPOSPHERE
	q_U	Free troposphere specific humidity	
	θ_U	Free troposphere potential temperature	
1 km	q_L	PBL specific humidity	PBL
	θ_L	PBL potential temperature	
	b	Vegetation cover	
10 cm	T_T	Surface layer temperature	SURFACE LAYER
	s_T	Surface layer relative humidity	
4 m	T_D	Deep soil layer temperature	DEEP SOIL LAYER
	s_D	Deep soil layer relative humidity	

An extension of D’Andrea, Provenzale, Vautard, de Noblet-Decoudré, GRL 2006, Baudena, D’Andrea, Provenzale, WRR 2008 and Baudena, Provenzale, HESS 2008

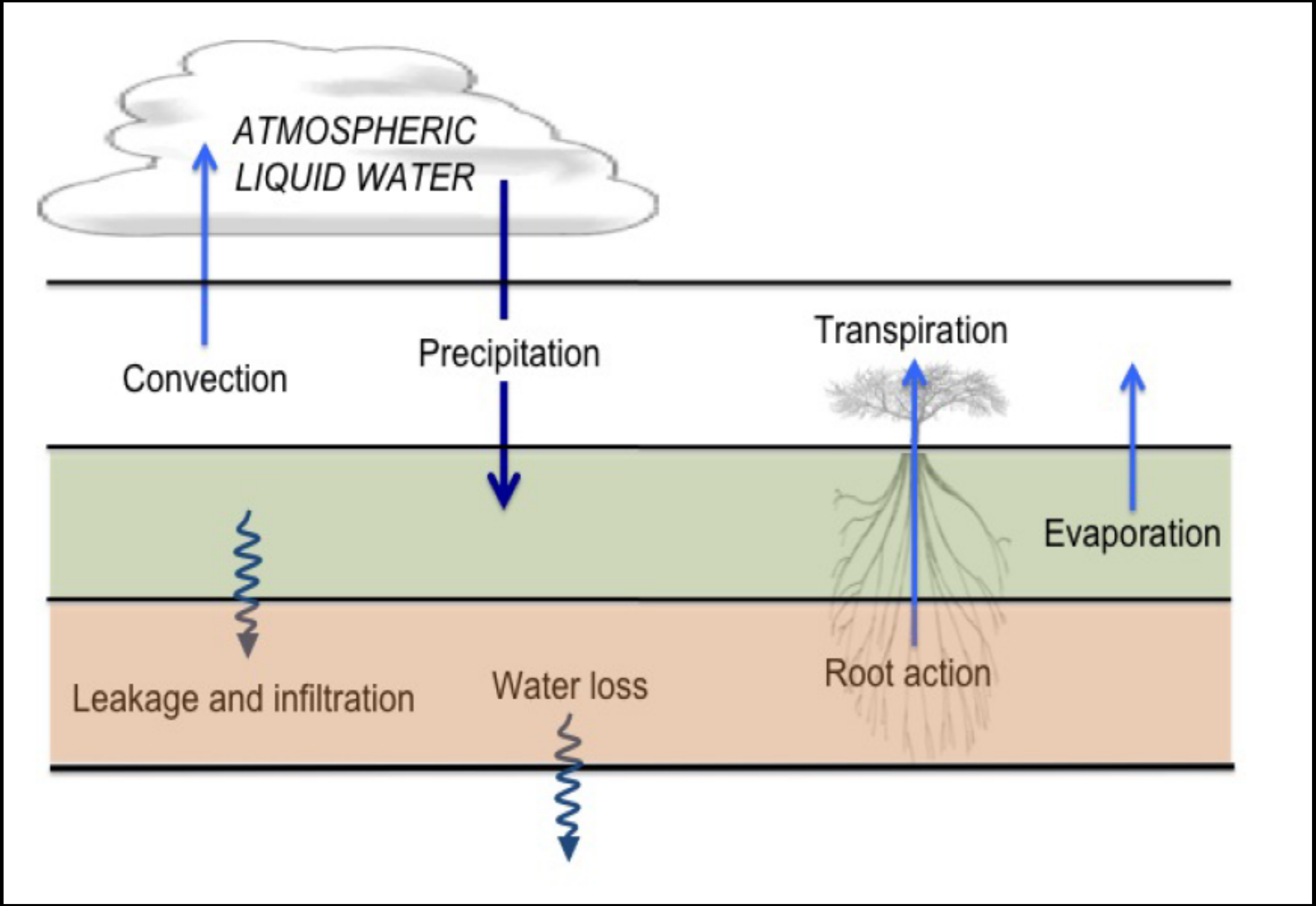
Model cycles and feedbacks

Heat



Model cycles and feedbacks

Moisture



Equations for the atmosphere

$$\rho_U h_U c_p \frac{\mathrm{d}\theta_U}{\mathrm{d}t} = L_e \frac{\Delta W}{\delta t} + \rho_L c_p h_L \frac{\widetilde{\Delta\theta}}{\delta t} + (1-\epsilon_L)\left(\epsilon_U + \epsilon_W \frac{W}{W_0}\right)\epsilon_S \sigma T_T^4 + \left(\epsilon_U + \epsilon_W \frac{W}{W_0}\right)\epsilon_L \sigma T_L^4 - 2\left(\epsilon_U + \epsilon_W \frac{W}{W_0}\right)\sigma T_U^4 \tag{1}$$

$$\rho_L h_L c_p \frac{\mathrm{d}\theta_L}{\mathrm{d}t} = Q_s - \rho_L c_p h_L \frac{\widetilde{\Delta\theta}}{\delta t} + \epsilon_L \epsilon_S \sigma T_T^4 + \left(\epsilon_U + \epsilon_W \frac{W}{W_0}\right)\epsilon_L \sigma T_U^4 - 2\epsilon_L \sigma T_L^4 \tag{2}$$

$$\rho_U h_U \frac{\mathrm{d}q_U}{\mathrm{d}t} = \rho_L h_L \frac{\widetilde{\Delta q}}{\delta t} - \frac{\Delta W}{\delta t} \tag{3}$$

$$\rho_L h_L \frac{\mathrm{d}q_L}{\mathrm{d}t} = \left[(1-b) + Sb\right] E\left(s_T, q_L\right) + b\, R\left(s_D, q_L\right) - \rho_L h_L \frac{\widetilde{\Delta q}}{\delta t} \tag{4}$$

$$\frac{\mathrm{d}W}{\mathrm{d}t} = \frac{\Delta W}{\delta t} - P \tag{5}$$

$$\Delta W = \rho h_U \left[q_U - q_{sat}(T_U)\right]$$

$$P = \begin{cases} (f_c + f_W)\frac{W}{\delta t} & \text{if } f_c + f_W \leqslant 1 \\ \frac{W}{\delta t} & \text{if } f_c + f_W > 1 \end{cases}$$

Equations for the soil

$$
\begin{aligned}
\rho_s c_{ps} Z_T \frac{\mathrm{d}T_T}{\mathrm{d}t} = {} & \{1-[b\alpha_v+(1-b)\alpha_e]\}F_{rad}(W) - Q_s \\
& - \epsilon_S \sigma T_T^4 + \epsilon_L \sigma T_L^4 \\
& + (1-\epsilon_L)\,\epsilon_S\left(\epsilon_U + \epsilon_W \frac{W}{W_0}\right)\sigma T_U^4 \\
& - L_e E\,(s_T, q_L)\,[(1-b)+Sb] \\
& - L_e b R\,(s_D, q_L) \\
& - \rho_s c_{ps}(Z_T+Z_D)\frac{T_T-T_D}{\tau_T}
\end{aligned}
\tag{7}
$$

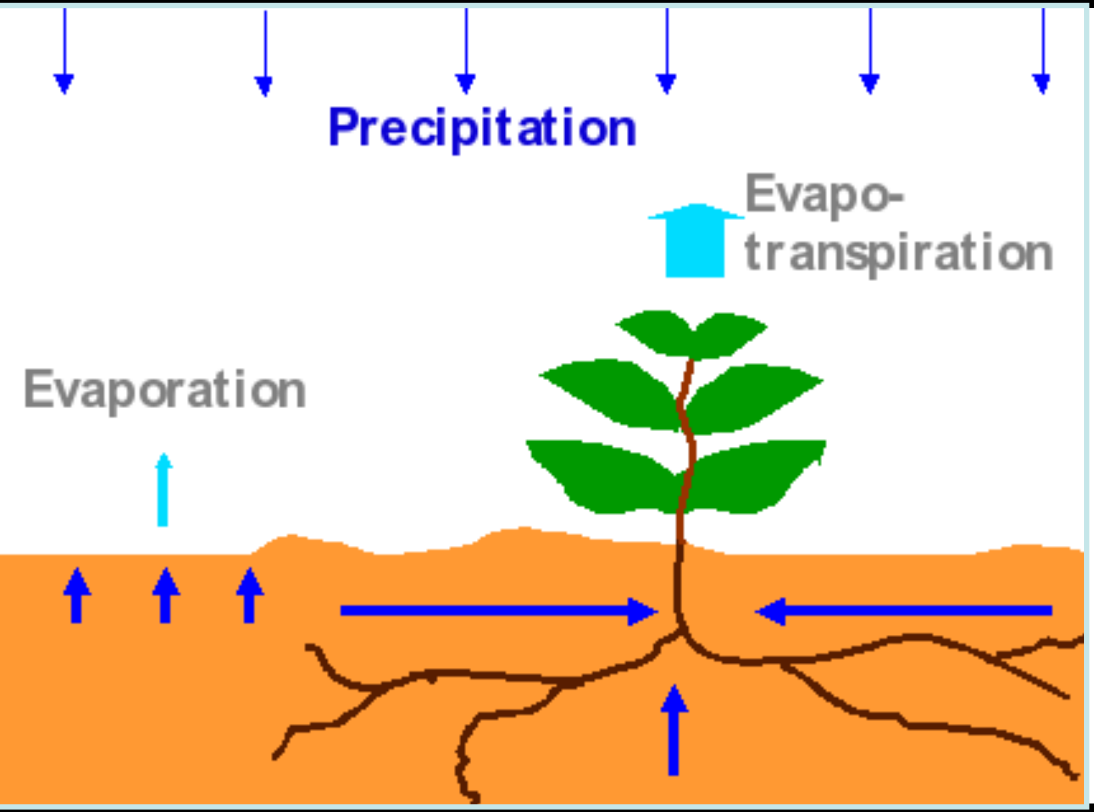
$$
Z_D \frac{\mathrm{d}T_D}{\mathrm{d}t} = -(Z_T+Z_D)\frac{T_D-T_T}{\tau_T} - (Z_D+Z_0)\frac{T_D-T_0}{\tau_D} \tag{8}
$$

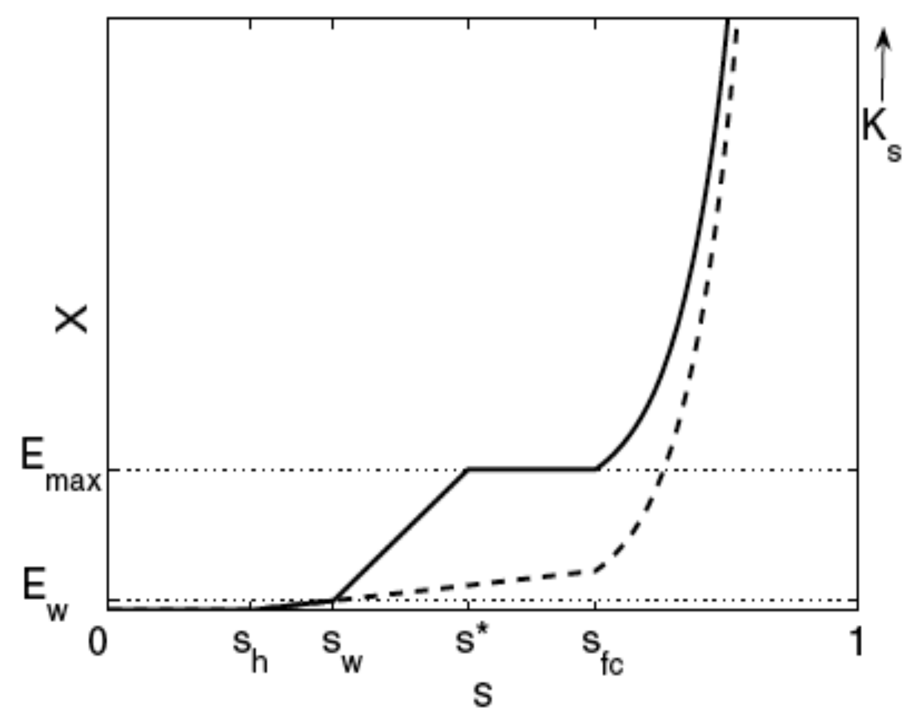
$$
\begin{aligned}
\rho_W n Z_T \frac{\mathrm{d}s_T}{\mathrm{d}t} = {} & P - E\,(s_T, q_L)\,[(1-b)+Sb] - L(s_T) \\
& - \rho_W n Z_T \frac{\Delta I\,(s_T)}{\delta t}
\end{aligned}
\tag{9}
$$

$$
\begin{aligned}
\rho_W n Z_D \frac{\mathrm{d}s_D}{\mathrm{d}t} = {} & \rho_W n Z_T \frac{\Delta I\,(s_T)}{\delta t} + L(s_T) - bR\,(s_D, q_L) \\
& - L(s_D) - \rho_W n Z_D \frac{\Delta I\,(s_D)}{\delta t}
\end{aligned}
\tag{10}
$$

$$
\frac{\mathrm{d}b}{\mathrm{d}t} = g(s_T, s_D)b\,(1-b) - \mu(s_D)b \tag{11}
$$

Evapotranspiration





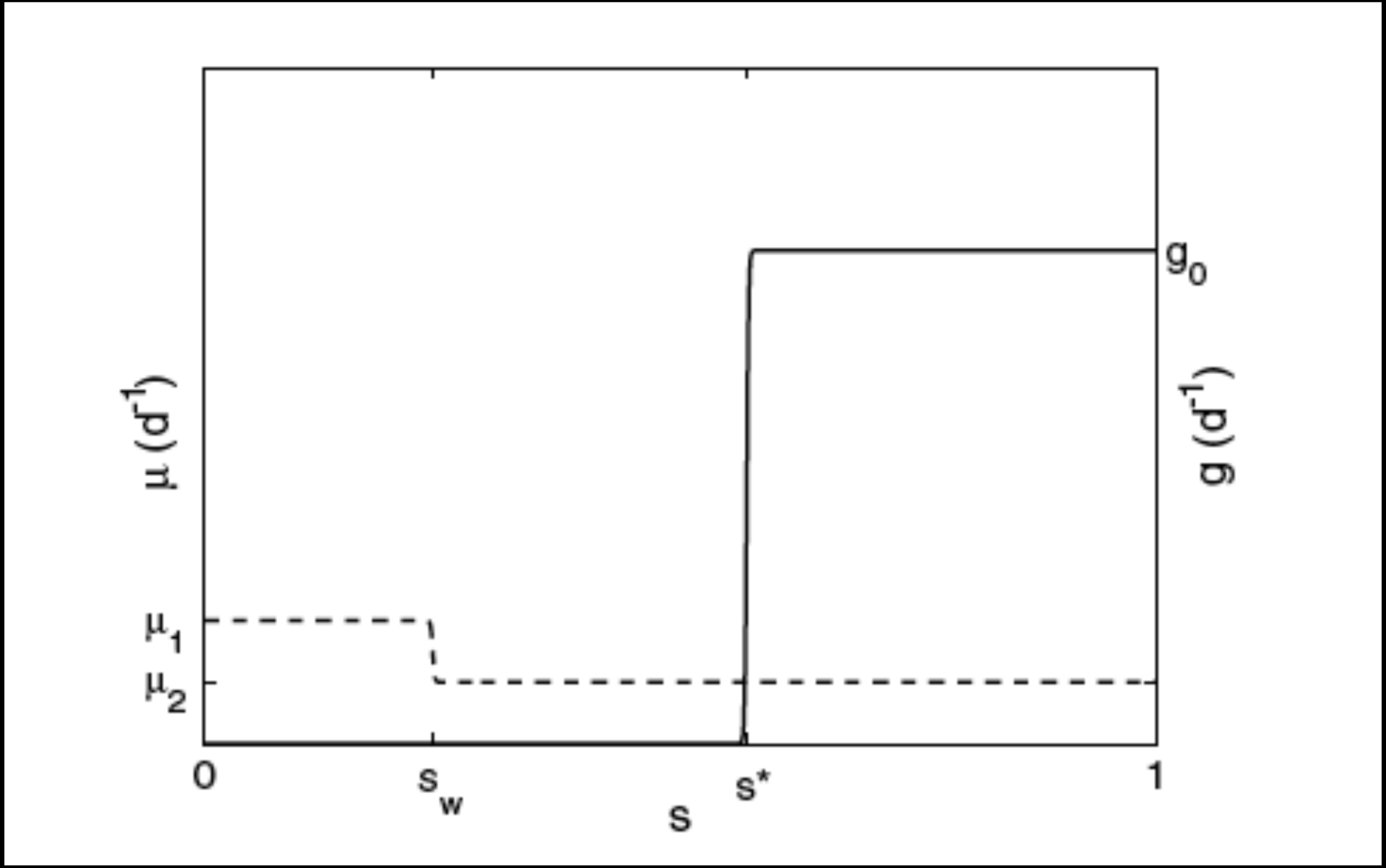
$$X(s,b) = E + L = b\chi_b(q_s) + (1-b)\chi_0(q_s)$$

Baudena, D' Andrea, AP, WRR, 2009

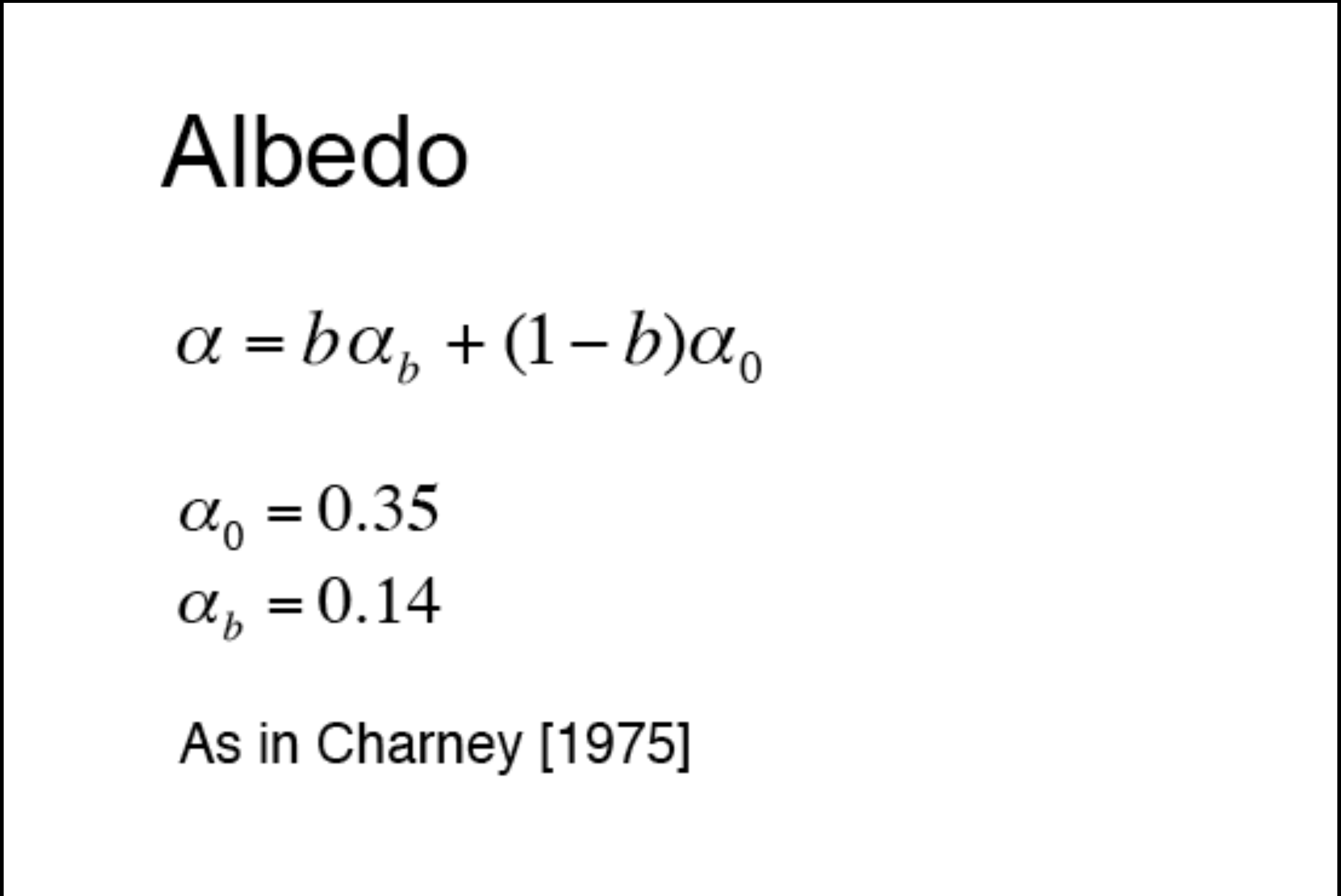
Vegetation dynamics

Levins, *Bull. Entomol. Soc. Am.* 1969; Tilman, *Ecology* 1994

$$\frac{\mathrm{d}b}{\mathrm{d}t} = gb(1-b) - \mu b.$$



Baudena, AP, HESS 2008; Baudena, D’Andrea, AP, WRR 2009



Albedo

$$\alpha = b\alpha_b + (1-b)\alpha_0$$

$$\alpha_0 = 0.35$$

$$\alpha_b = 0.14$$

As in Charney [1975]

Baudena, D’Andrea, AP, WRR 2009

Convection parameterization:

If $\theta_e = \theta_a \exp\frac{L_e q_a}{c_p \theta_a} > \theta_e^*$ convection occurs

We assume that convection is instantaneous

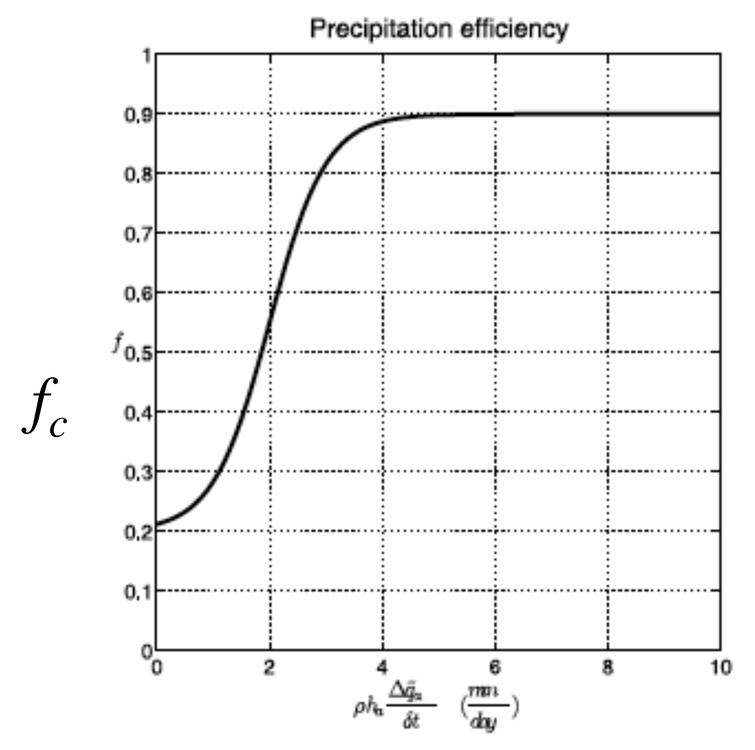
$$(\theta_L - \widetilde{\Delta\theta})e^{\frac{L_e(q_L - \widetilde{\Delta q})}{c_p(\theta_L - \widetilde{\Delta\theta})}} = (\theta_U + \widetilde{\Delta\theta})e^{\frac{L_e\left(q_U + \widetilde{\Delta q}\frac{\rho_L h_L}{\rho_U h_U}\right)}{c_p(\theta_U + \widetilde{\Delta\theta})}}.$$

$$\beta = \frac{c_p\widetilde{\Delta\theta}}{L_e\widetilde{\Delta q}}.$$

Precipitation: the fraction of liquid water which does not stay suspended

$$P = \begin{cases} (f_c + fw)\frac{W}{\delta t} & \text{if } f_c + fw \leqslant 1 \\ \frac{W}{\delta t} & \text{if } f_c + fw > 1 \end{cases}$$

$$fw = \frac{W}{W_0}$$



Results

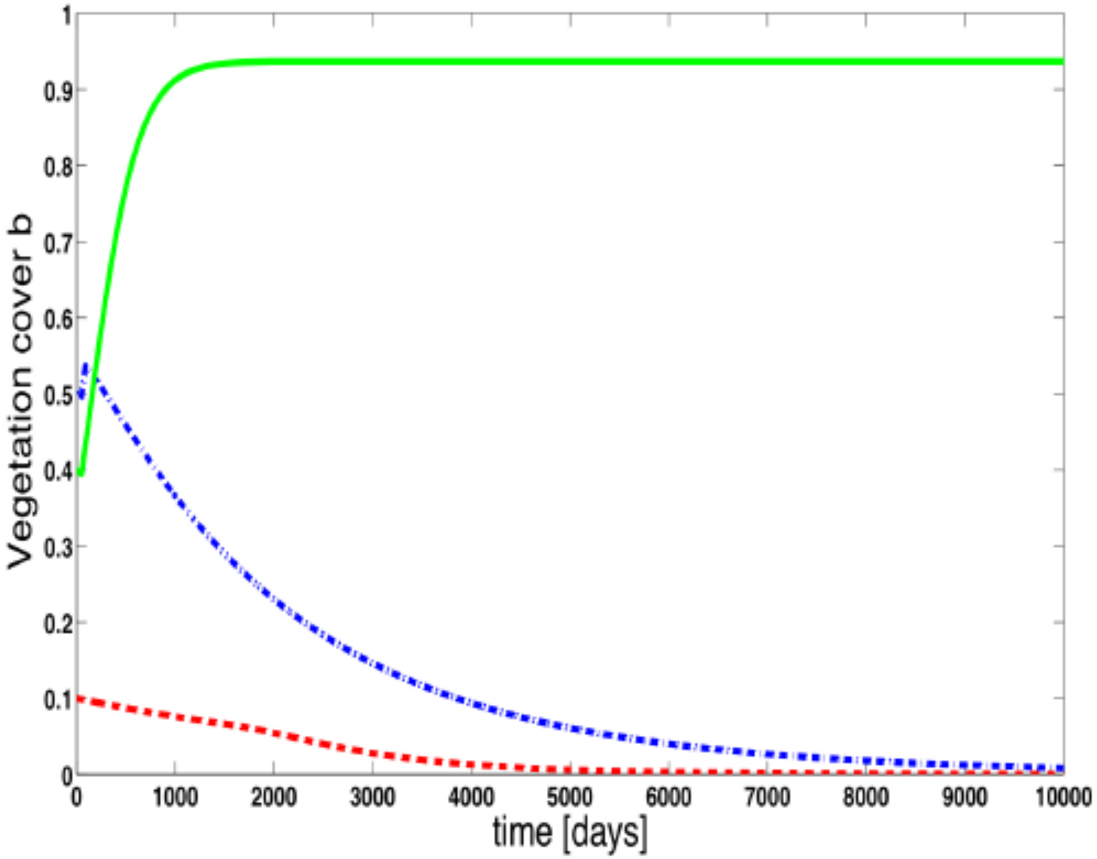


Figure 5. Vegetation cover temporal evolution starting from different initial conditions, and reaching the three different states: $\mathcal{D}$ (red dashed line), $\mathcal{F}$ (green continuous line), and $\mathcal{E}$ (blue dotted line). Note the different timescales needed to reach each of the states. For initial conditions value of deep soil water and vegetation cover, see Fig. 4.

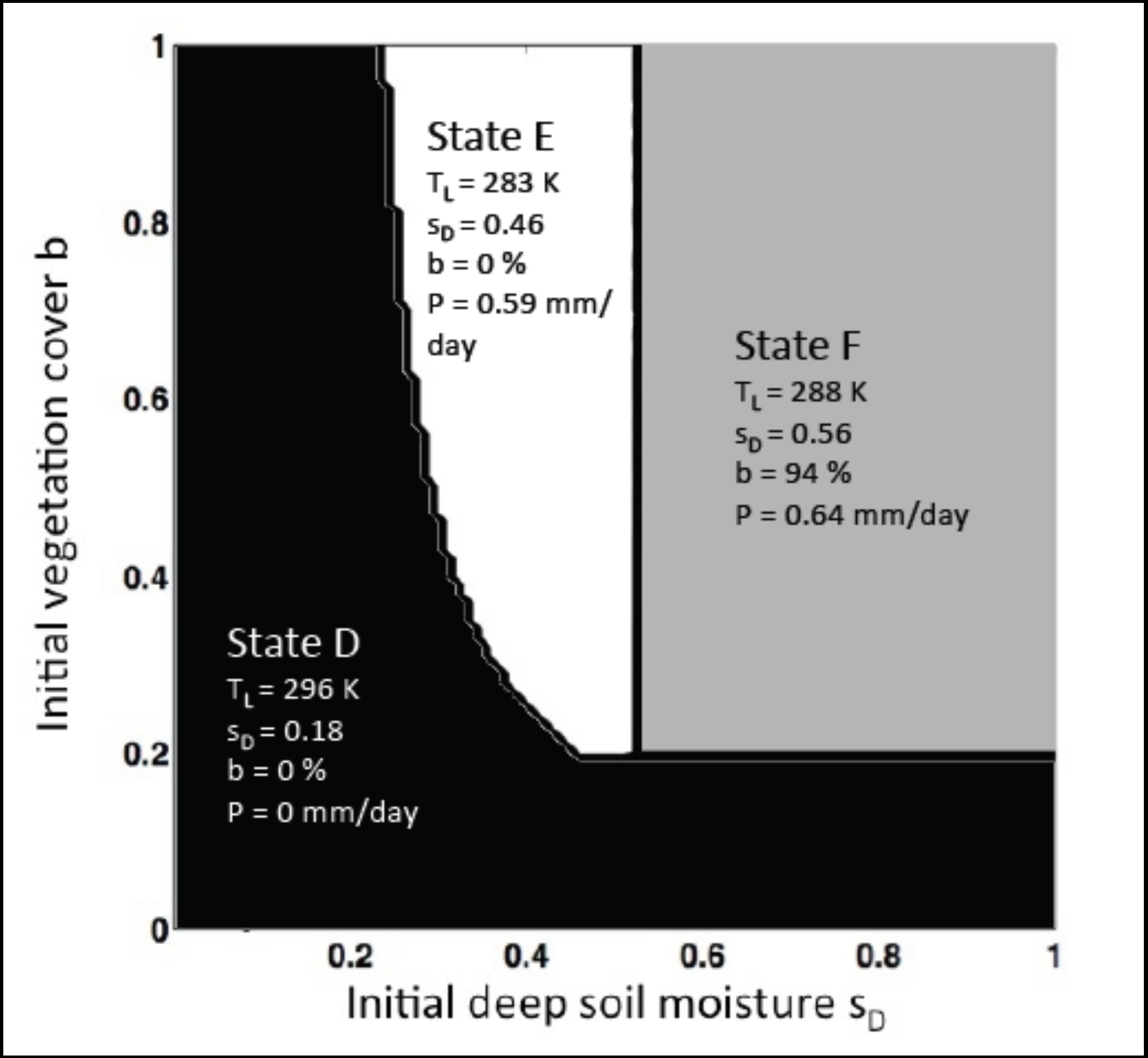


Figure 3. Equilibrium potential temperature in the PBL for the parameter values indicated in the Appendix, as a function of the initial conditions on s_D (x-axis) and on b (y-axis). The black area indicates the equilibrium state $\mathcal{D}$ (dry and hot, $\theta_L \simeq 23^\circ$ C), the white area indicates state $\mathcal{E}$ (wet and temperate, $\theta_L \simeq 9.6^\circ$ C), and the grey area indicate state $\mathcal{F}$ (dry and cold, $\theta_L \simeq 15.3^\circ$ C).

Conclusions from this simple model world

Transpiration from vegetation
is able to sustain a hydrologic cycle

For the same external forcing,
the model exhibits multiple steady states

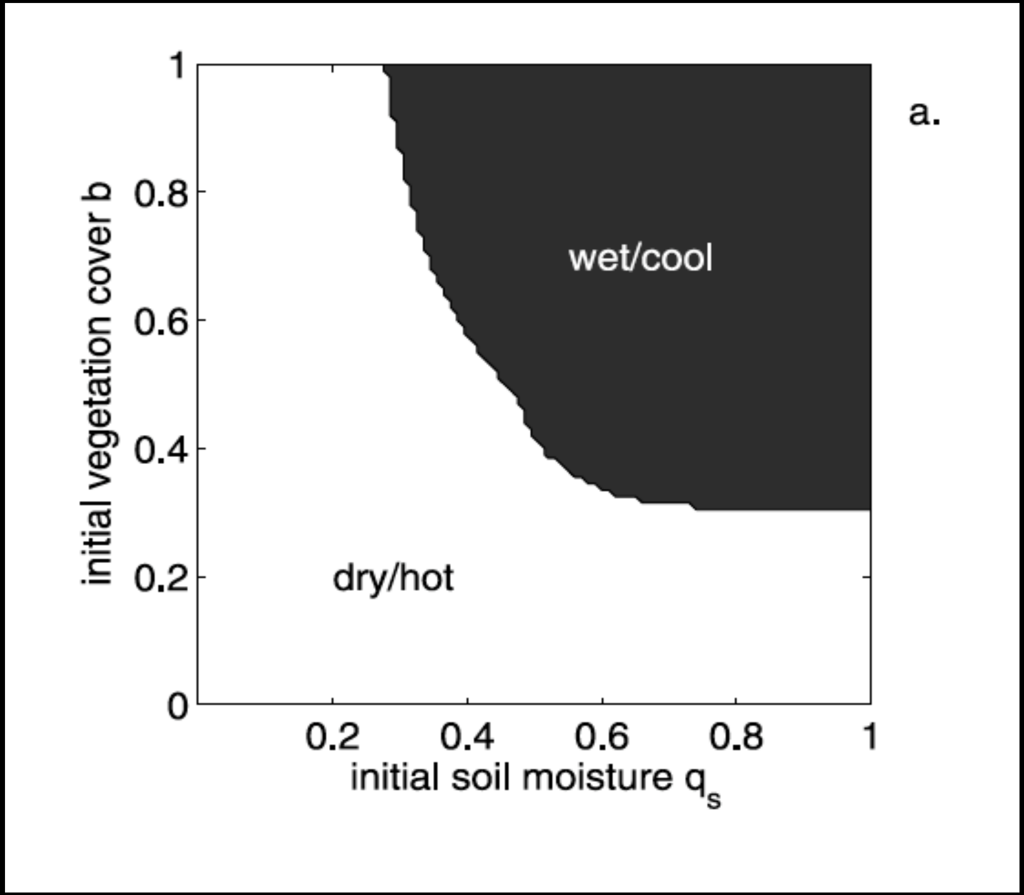
Transpiration feedback is more important
than albedo feedback

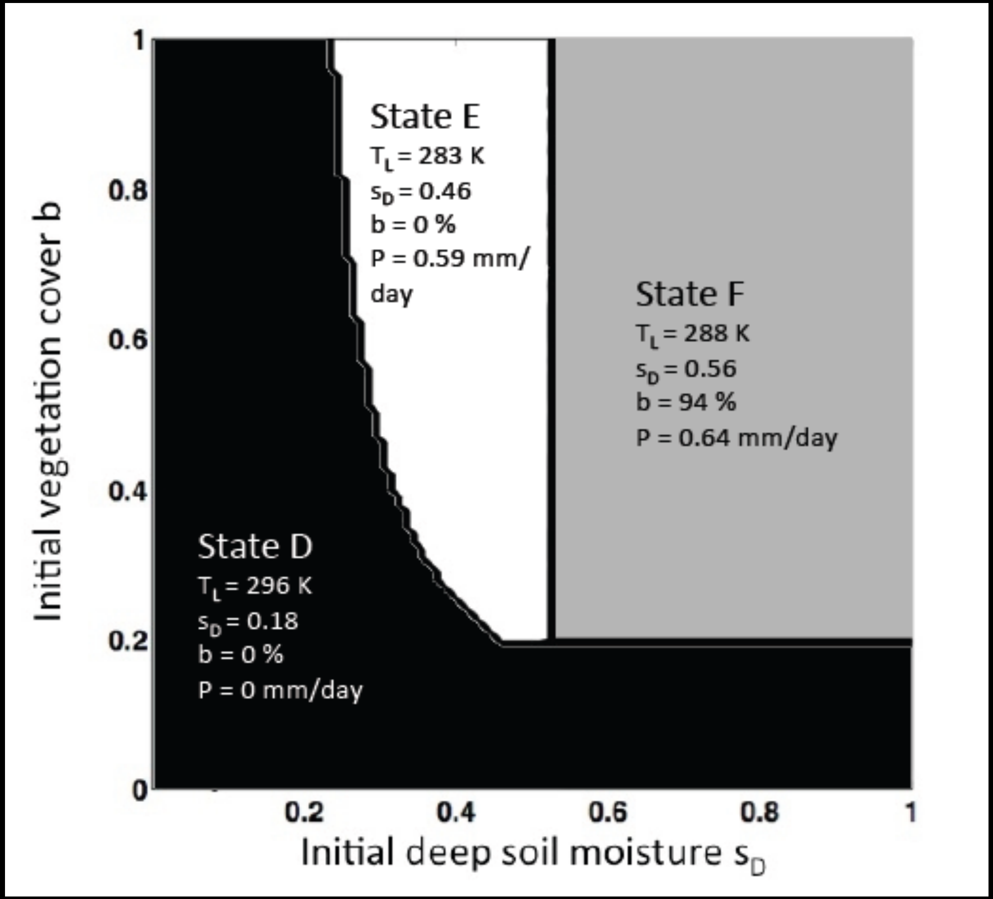
Importance of the convection parameterization

Remarks on feedbacks and upscaling

(perspective paper, Rietkerk et al 2011)

In many cases, local ecohydrological systems are characterized by multiple equilibria

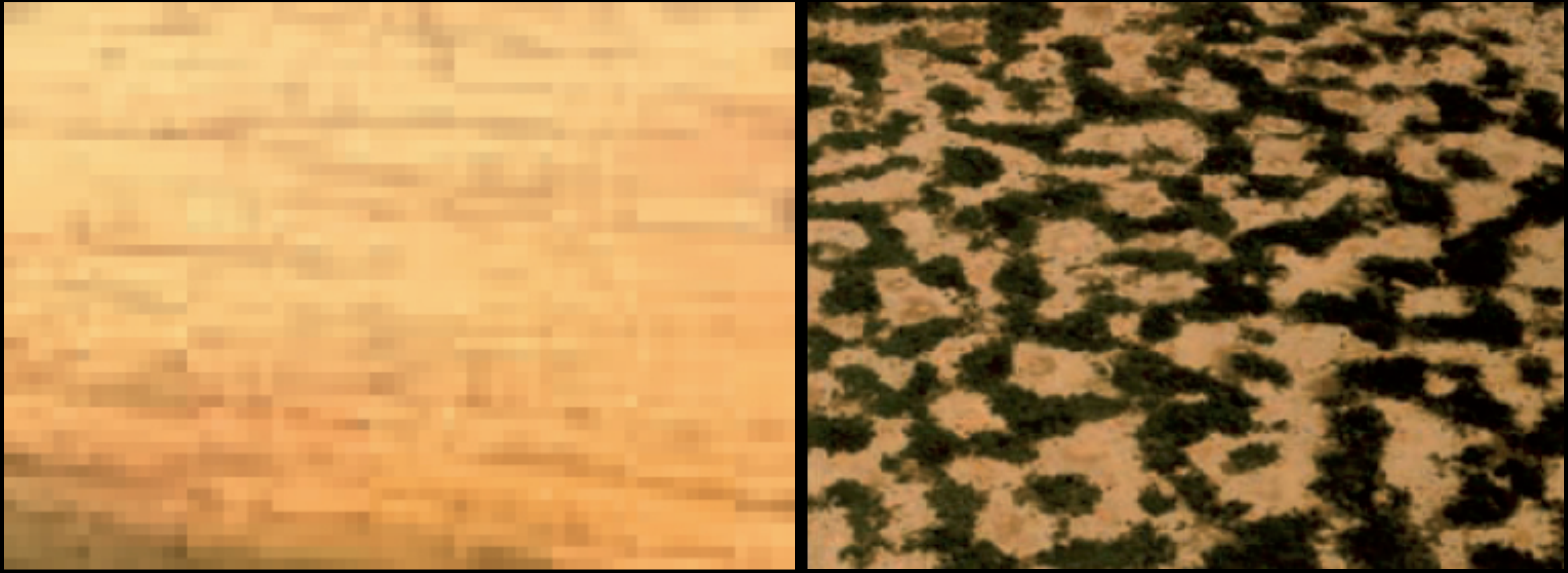




Remarks on feedbacks and upscaling

(perspective paper, Rietkerk et al 2011)

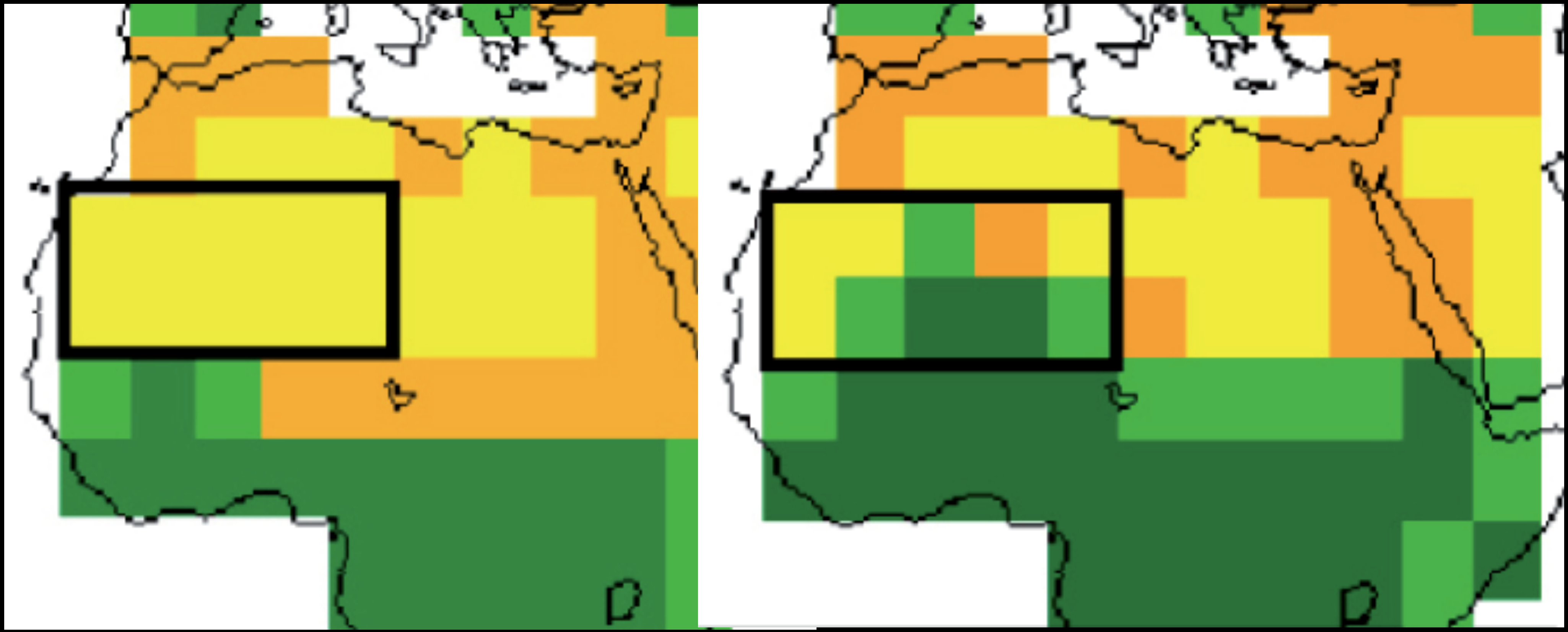
In many cases, local ecohydrological systems are characterized by multiple equilibria

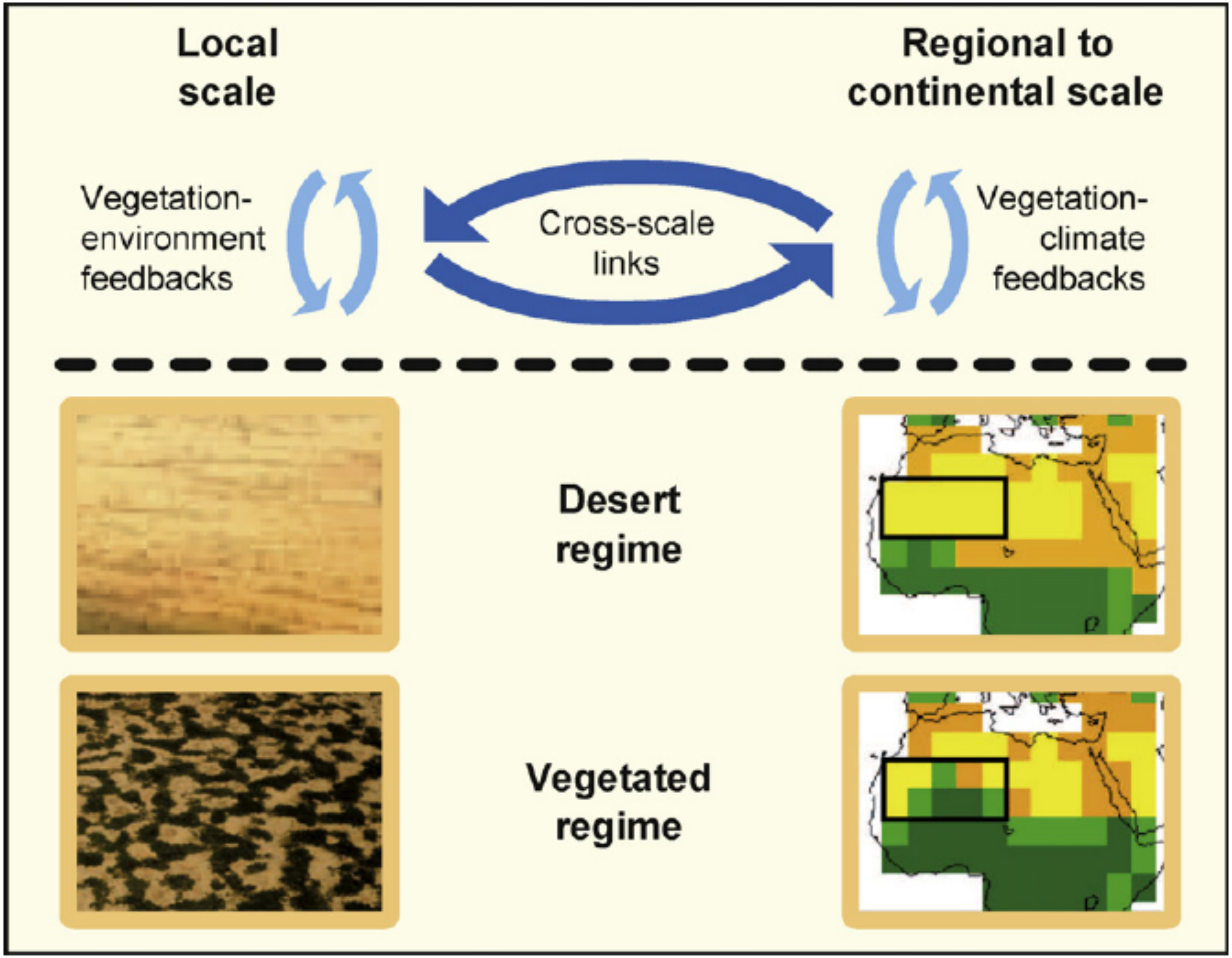


Remarks on feedbacks and upscaling

(Rietkerk et al 2009)

From local multiple equilibria
to regional and global climate ?





Local scale

Regional to continental scale

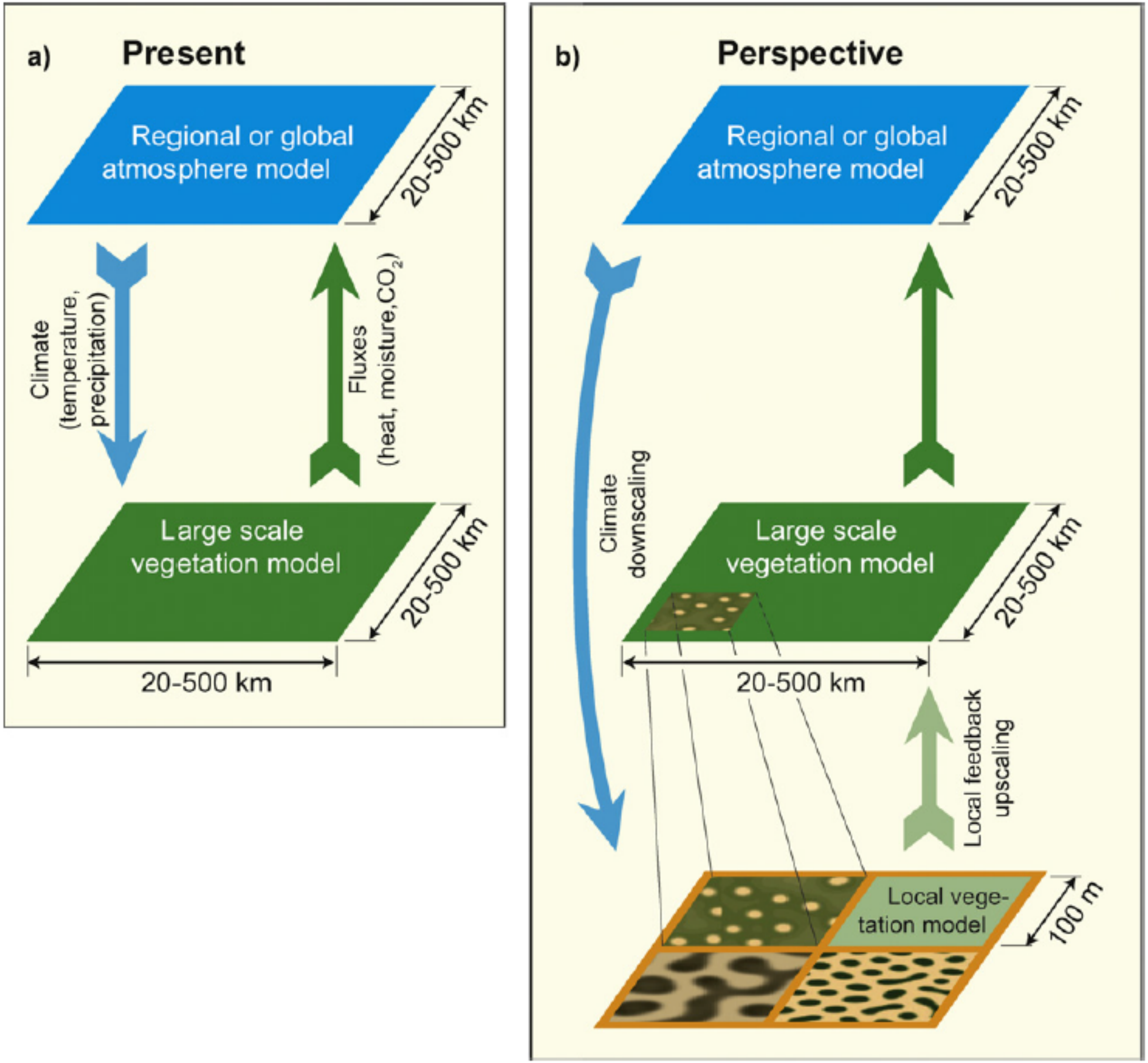
Vegetation-environment feedbacks

Cross-scale links

Vegetation-climate feedbacks

Desert regime

Vegetated regime


a)
Present
Regional or global atmosphere model
20-500 km
Climate (temperature, precipitation)
Fluxes (heat, moisture, CO$_2$)
Large scale vegetation model
20-500 km
20-500 km
b)
Perspective
Regional or global atmosphere model
20-500 km
Climate downscaling
Large scale vegetation model
20-500 km
20-500 km
Local feedback upscaling
Local vegetation model
100 m

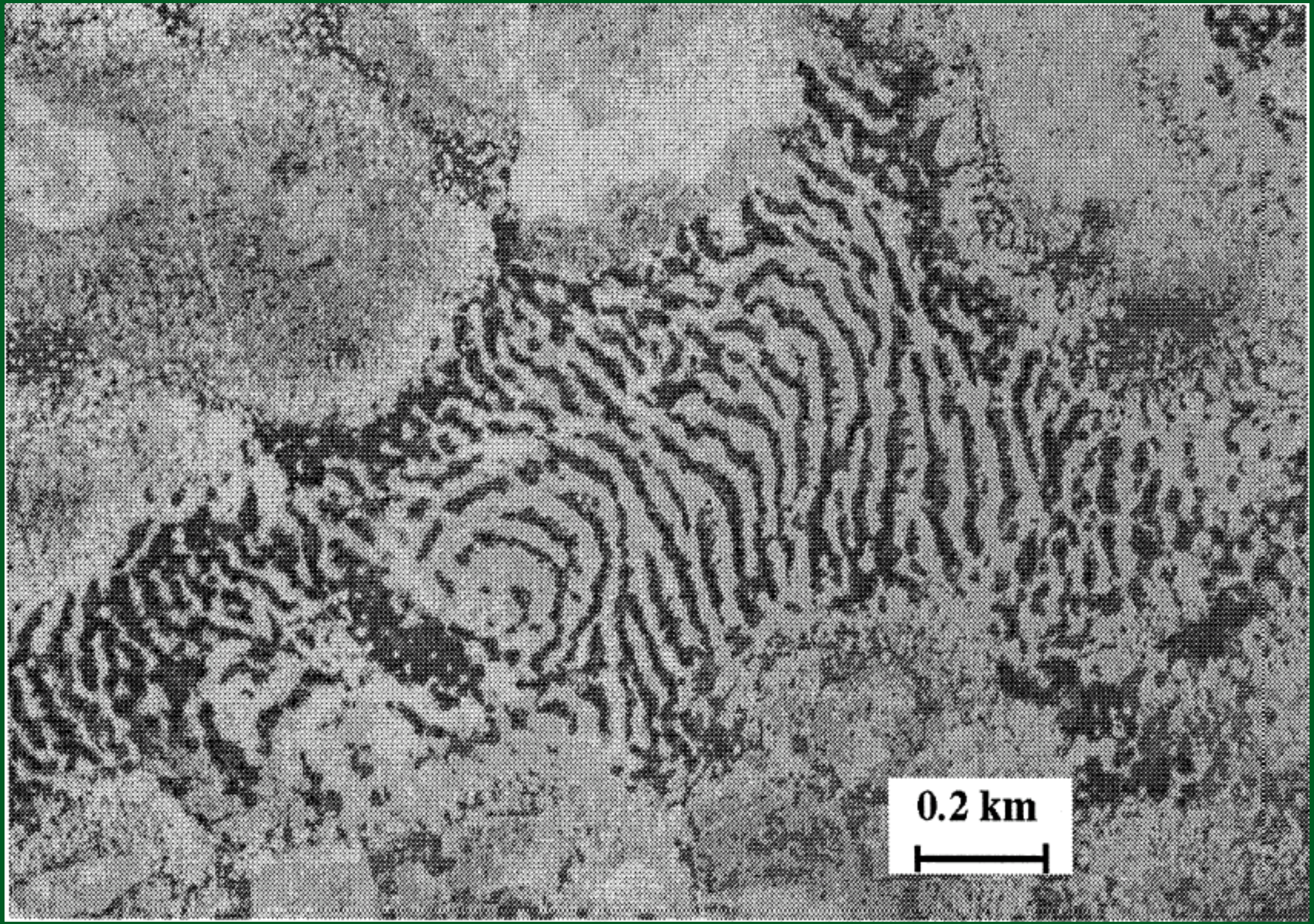
Vegetation patterns and evapotranspiration fluxes

M. Baudena, J. von Hardenberg, A. Provenzale, *AWR* 2013

Rietkerk et al., *The American Naturalist* **160** (4), 2002

In arid and semi-arid regions vegetation often forms patterned states

Vegetation patterns at landscape scale

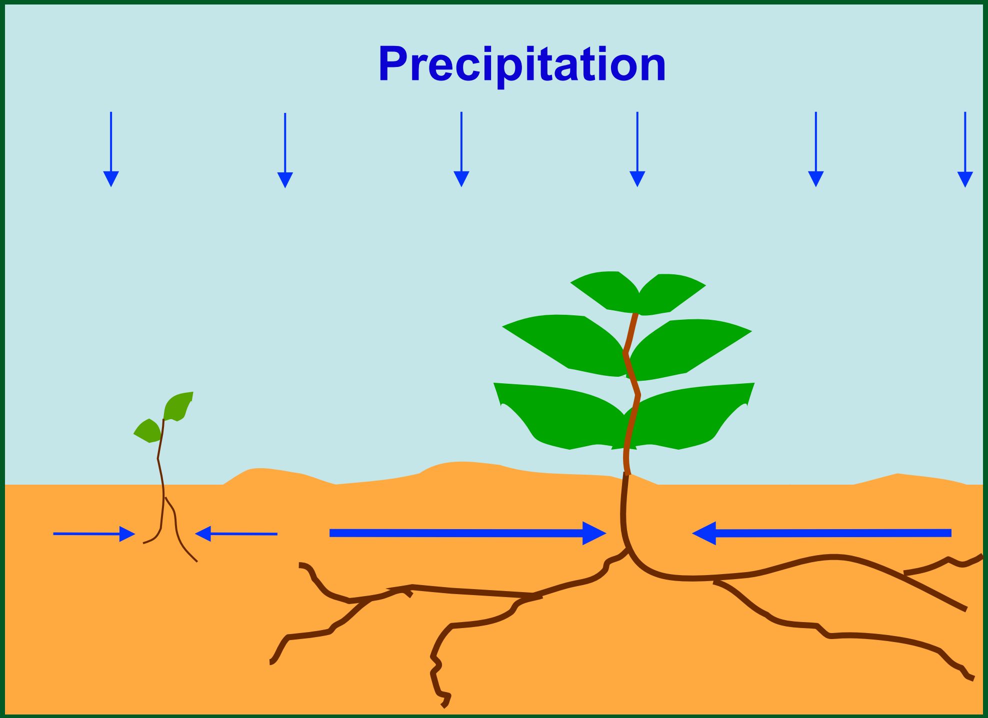


Valentin et al., *Catena* **37**, *1-24* (1999)

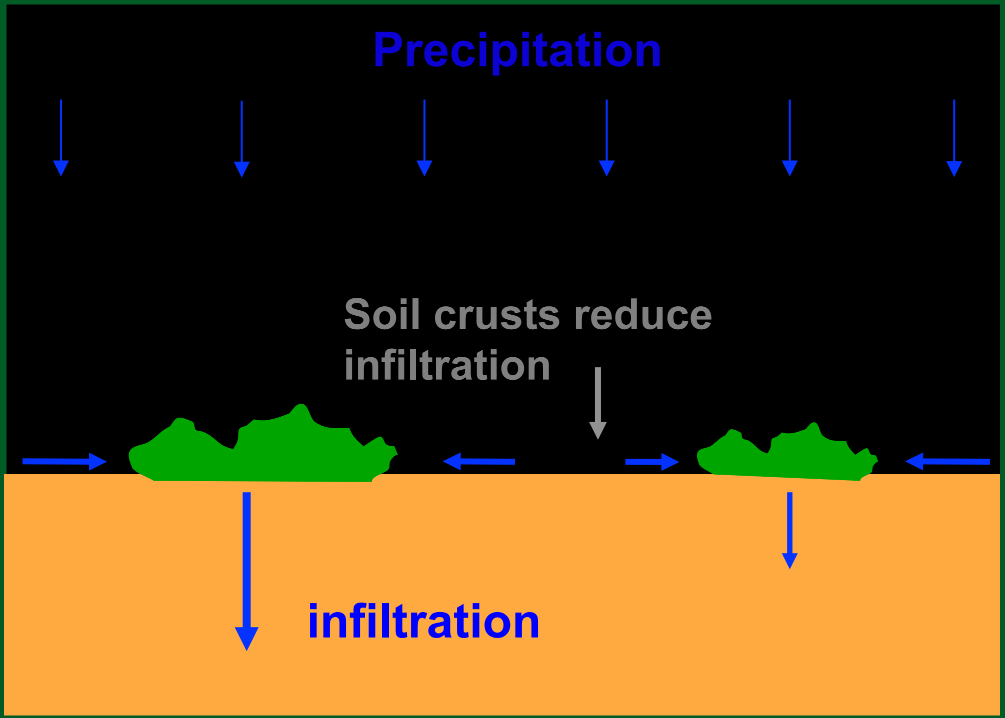
Feedbacks leading to vegetation patterns

Positive feedback between biomass and water + competition

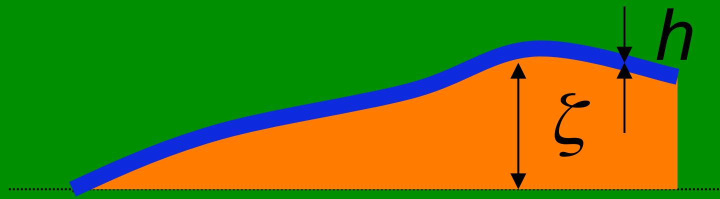
Water uptake by roots



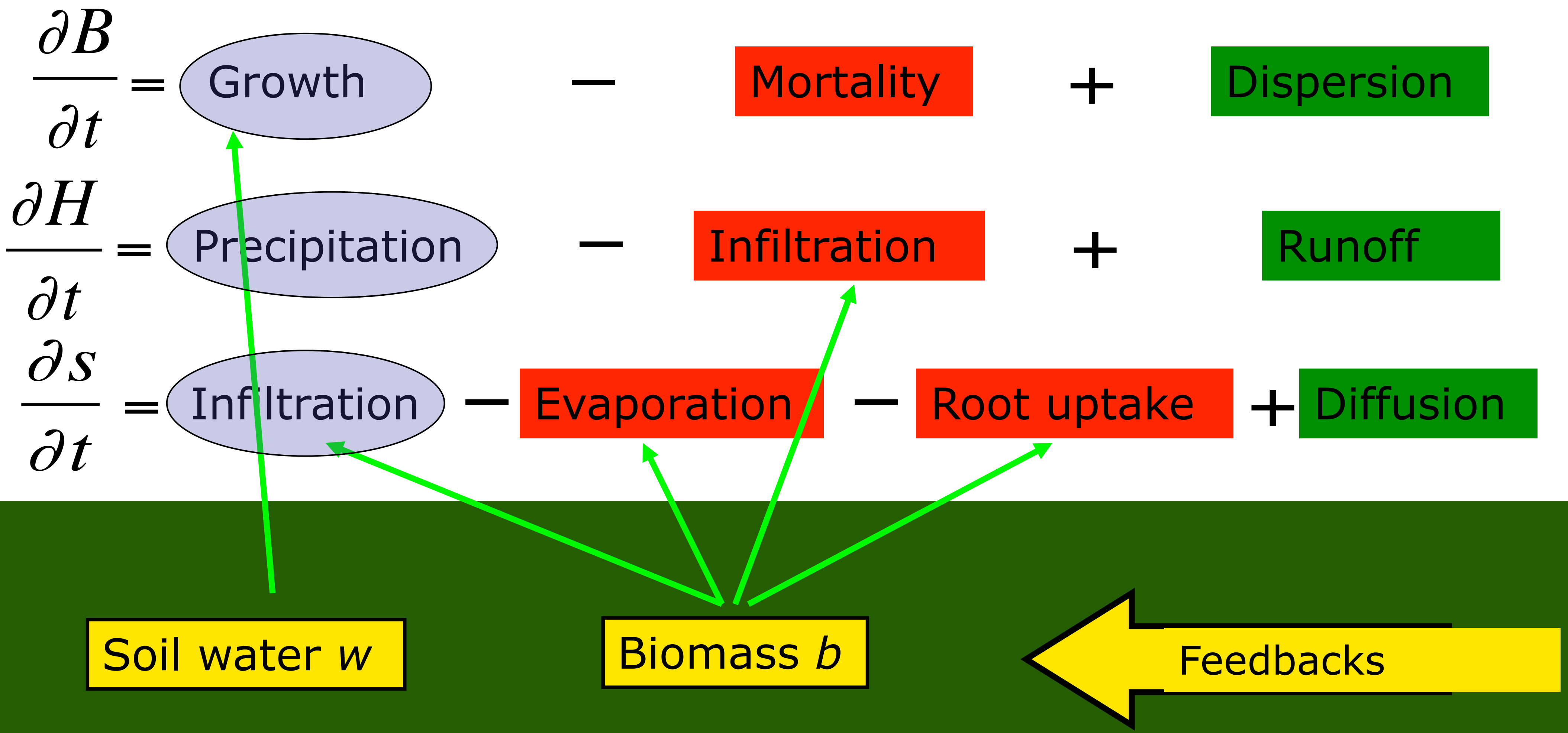
Increased infiltration



Vegetation - soil moisture - surface flow model



Plant biomass density $b(\mathbf{x},t)$ [Kg/m^2]
Relative soil moisture $s(\mathbf{x},t)$
Surface water height $h(\mathbf{x},t)$ [mm] or [Kg/m^2]



$$\frac{\partial B}{\partial t} = \text{Growth} - \text{Mortality} + \text{Dispersion}$$

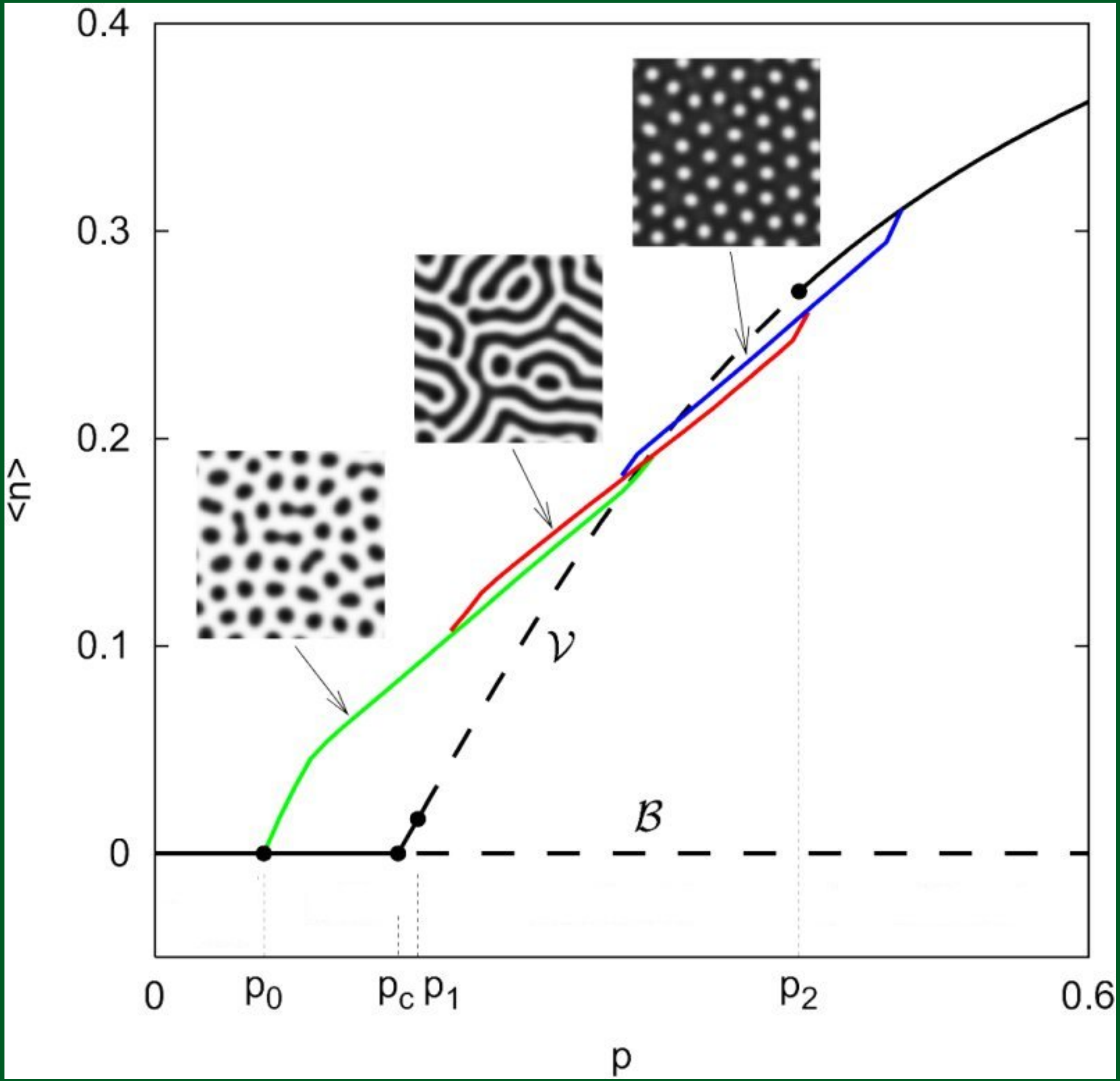
$$\frac{\partial H}{\partial t} = \text{Precipitation} - \text{Infiltration} + \text{Runoff}$$

$$\frac{\partial s}{\partial t} = \text{Infiltration} - \text{Evaporation} - \text{Root uptake} + \text{Diffusion}$$

Soil water *w*

Biomass *b*

Feedbacks



Vegetation patterns in arid and semi-arid regions

Gilad et al PRL 2004, JTB 2007, Kletter et al JTB 2009

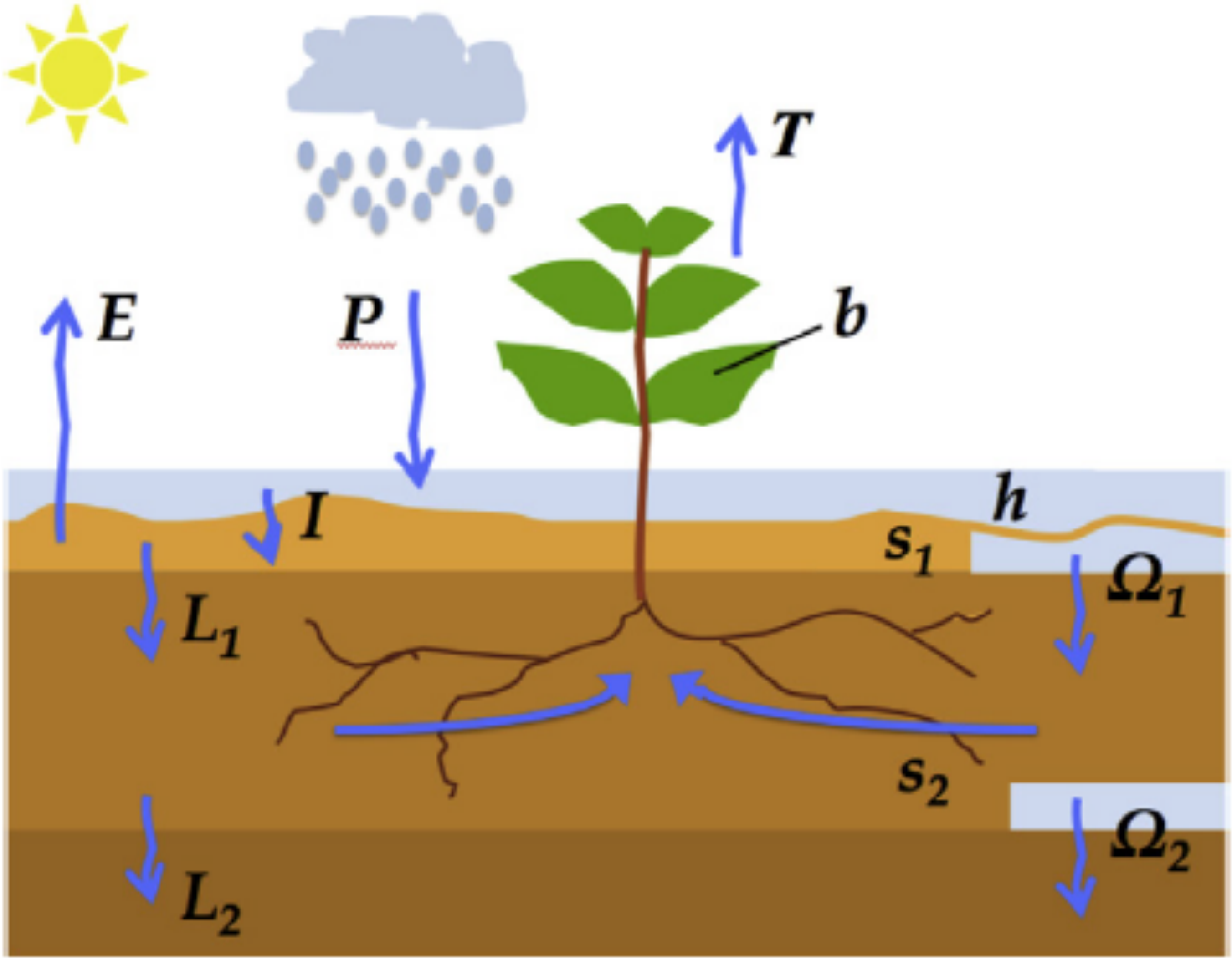


Fig. 1. Schematic representation of the model introduced here. The four prognostic variables are b (biomass density), s_1 surface soil moisture; s_2 deep soil moisture; h surface water. The water fluxes are: P, precipitation; E evaporation; T transpiration; I infiltration into the soil; $L_{1,2}$ leakage losses; $\Omega_{1,2}$ saturation excess water.

Vegetation - soil moisture - surface flow model

Plant biomass density $B(\mathbf{x},t)$ [Kg/m^2]
Relative soil moisture layer 1, $s_1(\mathbf{x},t)$
Relative soil moisture layer 2, $s_2(\mathbf{x},t)$
Surface water height $H(\mathbf{x},t)$ [mm] or [Kg/m^2]

$$\frac{\partial B}{\partial t} = \text{Growth} - \text{Mortality} + \text{Dispersion}$$

$$\cancel{\frac{\partial H}{\partial t}} = \text{Precipitation} - \text{Infiltration} + \text{Runoff}$$

$$\frac{\partial s_1}{\partial t} = \text{Infiltration} - \text{Evaporation} - \text{Infiltration/ Leakage to 2} + \cancel{\text{Diffusion}}$$

$$\frac{\partial s_2}{\partial t} = \text{Infiltration/ Leak from 1} - \text{Infiltration/ Leakage to D} - \text{Root uptake} + \cancel{\text{Diffusion}}$$

Results 1: Dependence of normalized evapotranspiration fluxes on the type of pattern

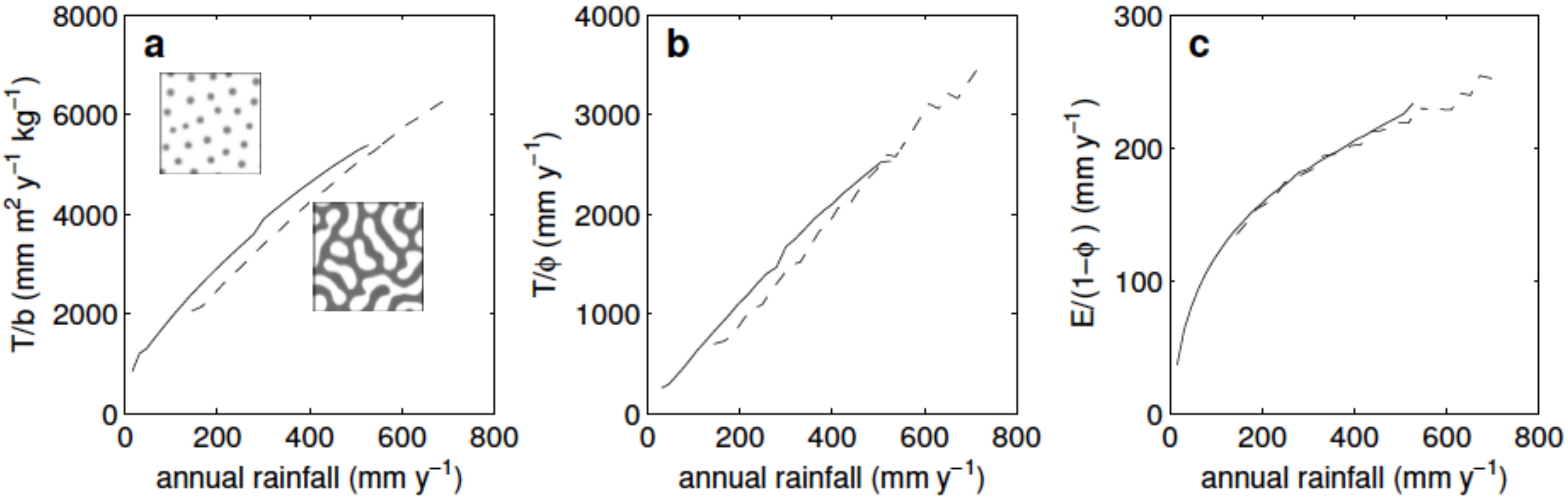


Figure 3: Evapotranspiration fluxes variation along a rainfall gradient. a. Transpiration flux T per unit biomass density b; b. transpiration flux T per unit vegetated area ϕ; c. evaporation flux E per unit bare soil $(1-\phi)$. Continuous line, spots; dashed line: stripes. Evaporation and transpiration fluxes are averaged in the five days after rainfall events, during the last 100 year of each run, and above the whole plot.

A difference of 10-15%. Is this much or not ?

Results 2: Dynamic vs fixed vegetation

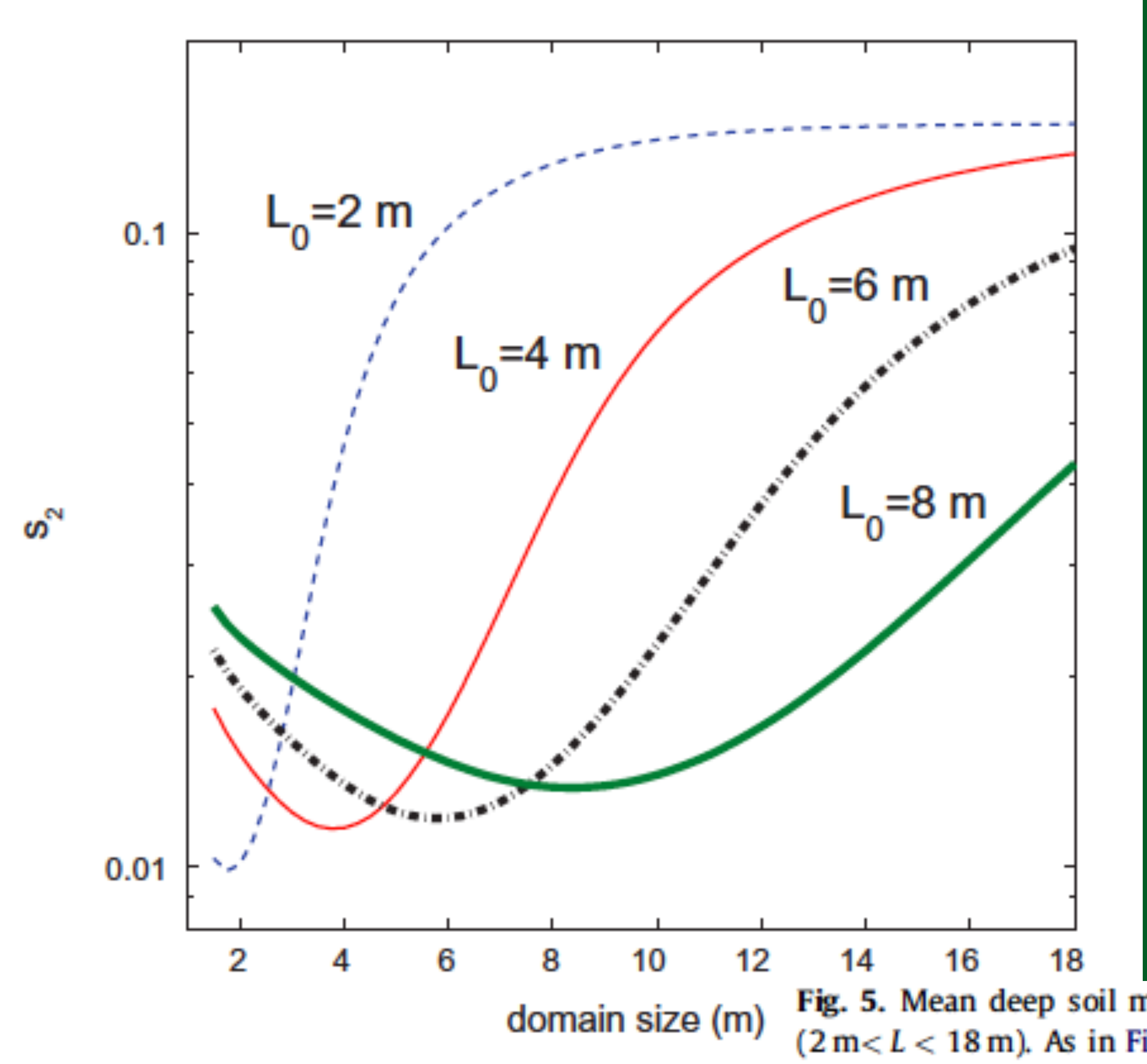


Fig. 5. Mean deep soil moisture (s_2) in bare soil, as a function of domain size L ($2\,\mathrm{m} < L < 18\,\mathrm{m}$). As in Fig. 4, vegetation is not dynamic. The different curves are obtained by rescaling different spot solutions originally obtained with a different value of L_0: blue dashed line, L_0=2 m; red continuous line, L_0=4 m; black dash-dotted line, L_0=6 m; green thick continuous line, L_0=8 m. The mean annual precipitation rate is 220 mm y^{-1}. All spatial averages are calculated for periods of five days after rainfall events, in the last 100 years of each simulation. In all cases, soil moisture in bare soil is minimum for the original, dynamically consistent solution. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

In the end:

Differences in the type of pattern generate differences in transpiration fluxes.
Between spots and strips, the difference is about 10-15% in the five days after rainfall events.

There is a minimum of soil water in bare soil
(a maximum in soil water uptake from bare soil)
for vegetation with self-consistent dynamics