Approaches to Modelling Evaporation in Climate Models

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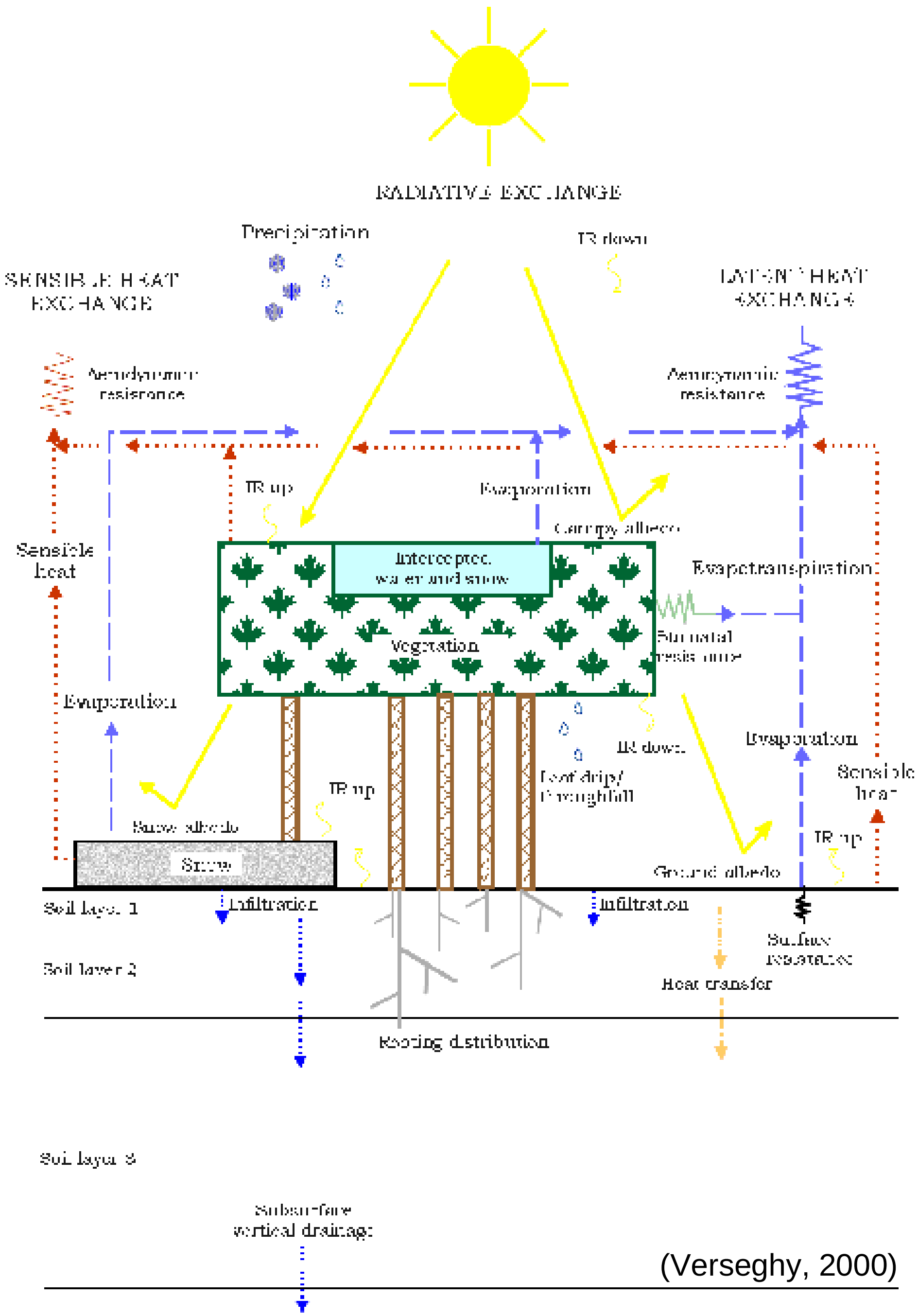


(Verseghy, 2000)

The Canadian Land Surface Scheme (CLASS)

Originally developed for the CGCM; treats fluxes of energy and water at the land surface

Thermally separate vegetation canopy, snow cover and three soil layers.

Four main vegetation structural types identified (needleleaf trees, broadleaf trees, crops and grass); parameters are aggregated at each time step to define representative canopy characteristics.

Up to four subareas allowed over each model grid cell: vegetation covered, bare soil, snow with vegetation and snow over bare soil.

One soil type for each grid cell.

Factors governing the flux of water vapour between the land surface and the atmosphere:

- **Available energy**
- **Available water**
- **Atmospheric state (e.g. near-surface stability, humidity)**
- **Vegetation physiological characteristics (e.g. stomatal conductance, stand architecture)**
- **Soil physical properties (e.g. soil moisture suction curve, pore volume)**

In numerical models, the standard approach to calculating the flux of water vapour between the earth's surface and the atmosphere is to use a one-dimensional "bulk transfer" formula derived from a simple flux-gradient relation and the standard logarithmic wind profile, as follows:

$$E = \rho_a c_D u_a \, [q_a - q_0]$$

where ρ_a is the density of the air, u_a and q_a the wind speed and specific humidity respectively of air at a given reference height, c_D the surface drag coefficient, and q_0 the specific humidity of air at the surface.

The drag coefficient under neutral atmospheric stability, c_{DN}, is a function of the reference height z_r, the zero-plane displacement height d, and the roughness length z_0:

$$c_{DN} = k^2/[\ \ln[(z_r - d)/z_0]^2$$

where k is the von Karman constant. The displacement height varies with vegetation height and stand architecture, and the roughness length varies with the surface irregularity. Standard empirical relations are used to calculate these parameters. Corrections are then applied to c_{DN} based on buoyancy effects caused by non-neutral stability conditions in the near-surface (“constant-flux”) layer of the atmosphere.

Returning to the “bulk transfer” equation for evaporation, the formulation presented above assumes a free liquid or frozen water surface, i.e. that the relative humidity at the surface is 100%. In the case of a soil surface under unsaturated conditions, the equation must in principle be modified by the inclusion of the surface relative humidity α:

$$E = \rho_a c_D u_a \left[q_a - \alpha q_{0,sat}(T_0)\right]$$

In practice, because of the strong variability of surface soil moisture, and the large uncertainties associated with identifying the relative humidity of a very thin surface layer, an empirical “beta” formulation is typically used in models, where β is a function of the bulk soil moisture:

$$E = \rho_a c_D u_a \beta \left[q_a - q_{0,sat}(T_0)\right]$$

a *and* b *coefficients for evaporation from the soil surface*

$$a = \exp\left[\frac{-gy_0}{R_w T_0}\right]$$

Philip (1957):
Requires extrapolation to surface soil moisture, to calculate ψ_0.

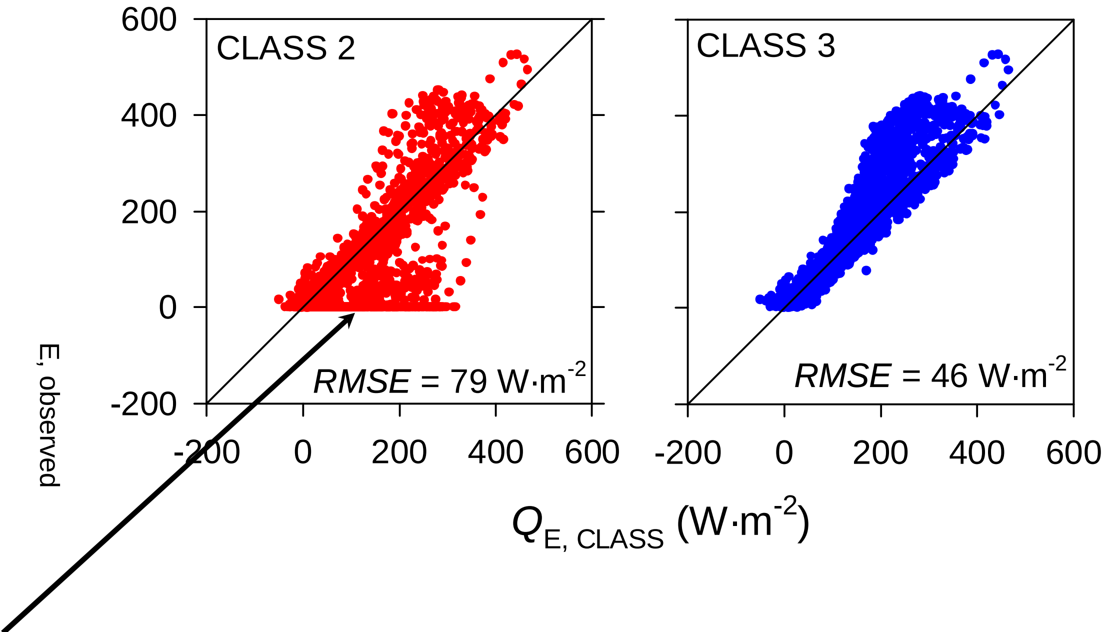
$$b = 1 \Big/ \left[1 + b_1 \left(q_{sat} - q\right)^{b_2}\right]$$

Wu et al. (2000):
Requires appropriate values for b_1 and b_2, soil moisture and porosity in top layer

$$b = \begin{cases} \frac{1}{4}\left[1 - \cos\left(\frac{q}{q_{fc}} p\right)\right]^2, & q_1 < q_{fc} \\ 1 & q_1 \geq q_{fc} \end{cases}$$

Lee and Pielke (1992):
Requires soil moisture in top layer, and the field capacity of the soil.

Evaporation from bare soil at Agassiz, British Columbia



- In CLASS 2 modelled evaporation from the soil surface suffers from periods of underestimation (including zero values), caused by an underestimation of moisture at the ground surface.
- In CLASS 3, the underestimation in modelled evaporation has been eliminated.

In the case of transpiration from a vegetation canopy, unless intercepted rain or snow is present, the stomatal resistance r_c must be taken into account: i.e. the resistance to movement of water vapour from the inside of the leaves to the outside, caused by the obstruction of the small leaf openings or stomata. This resistance, using an electrical analogy, is considered to operate in parallel with the atmospheric resistance r_a, which is determined by the wind speed and the drag coefficient. Defining $r_a = 1/(c_D u_a)$, the equation for water vapour flux from dry vegetation canopies can be written as:

$$E = \rho_a \,[q_a - q_{0,sat}(T_c)]/\,[r_a + r_c]$$

The stomatal resistance depends on ambient conditions such as incoming solar radiation, soil moisture, temperature, vapour pressure deficit, and plant physiology.

Canopy Conductance in CLASS 2.x

(Conductance, the inverse of resistance, tends to be more the standard usage in field studies)

- g_c scales linearly with leaf area index (Λ)

$$g_c(\Lambda) = g_{c,\max}\left[\Lambda / \Lambda_{\max}\right]$$

CLASS employs the multiplicative Jarvis-Stewart approach to represent the response to environmental stresses

$$g_c = \hat{g}_c\, f_1\left(K_\downarrow\right) f_2(\Delta e)\, f_3\left(y_{s,r}\right) f_4\left(T_a\right)$$

where:

$\hat{g}_c$ is a composite value of $g_c(\Lambda)$ over the 4 vegetation groups

$K_\downarrow$ is incoming solar radiation

Δe is vapour pressure deficit

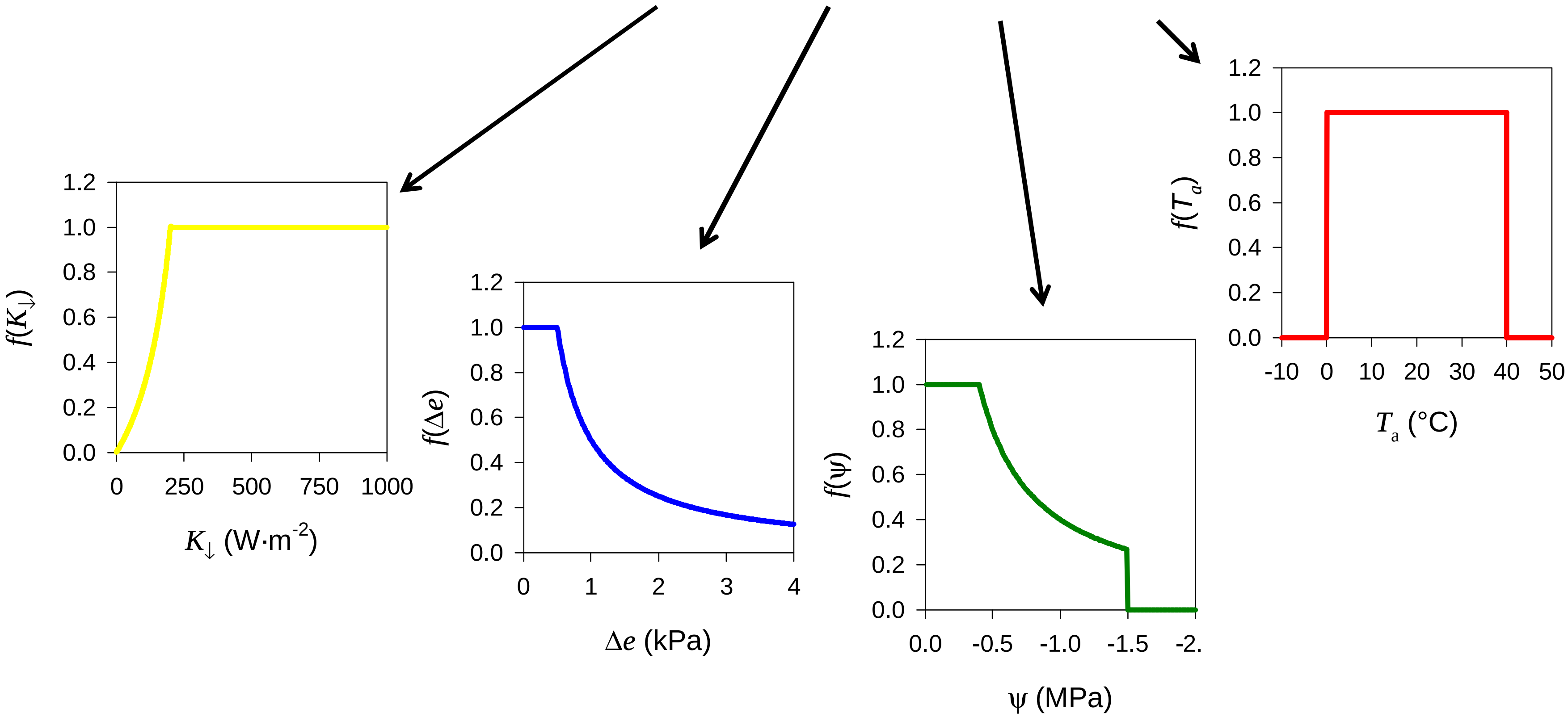
$\psi_{s,r}$ is soil water suction in the rooting zone

T_a is air temperature

Canopy Conductance in CLASS 2.x

- $g_{c,\max}$ is hard coded as 20 mm·s^{-1} for all vegetation types
- functions f_1 - f_4 are the same for all vegetation types

$$g_c = \hat{g}_c\, f_1(K_\downarrow)\, f_2(\Delta e)\, f_3(\psi_{s,r})\, f_4(T_a)$$

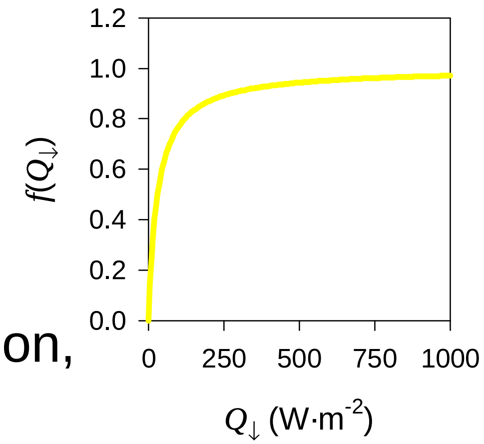


Canopy Conductance in CLASS 3.0

Following papers by Schulze, Kelliher, Leuning and Raupach (1995):

- The maximum unstressed stomatal conductance is $g_{s,\max}$
- We model a hyperbolic response to solar radiation

$$g_s = \frac{g_{s,\max} Q_{\downarrow}}{Q_{\downarrow} + Q_{l1/2}}$$



where:
$Q_{\downarrow}$ is incoming photosynthetically active radiation,
$Q_{l1/2}$ is the value of $Q_{\downarrow}$ where $g_s = g_{s,\max}/2$

Canopy Conductance in CLASS 3.0

- Assuming photosynthetically active radiation at height h (Q_h) declines exponentially through the canopy with cumulative leaf area index (x)

$$Q_h = Q_{\downarrow} \exp\left(-c_Q x\right)$$

where c_Q is an extinction coefficient.

- Differentiating with respect to x and assuming g_c is the parallel sum of g_s through the canopy, we can combine the previous two equations to yield canopy conductance in the absence of stress caused by humidity, water availability and air temperature, as

$$g_c = \frac{g_{s,\max}}{c_Q} \ln\left[\frac{Q_{\downarrow} + Q_{1/2}}{Q_{\downarrow} \exp\left(-c_Q \Lambda\right) + Q_{1/2}}\right]$$

where $Q_{1/2} = Q_{l\,1/2}/c_Q$.

Canopy Conductance in CLASS 3.0

- We can represent stress caused by humidity, water availability and temperature using multiplicative functions, as

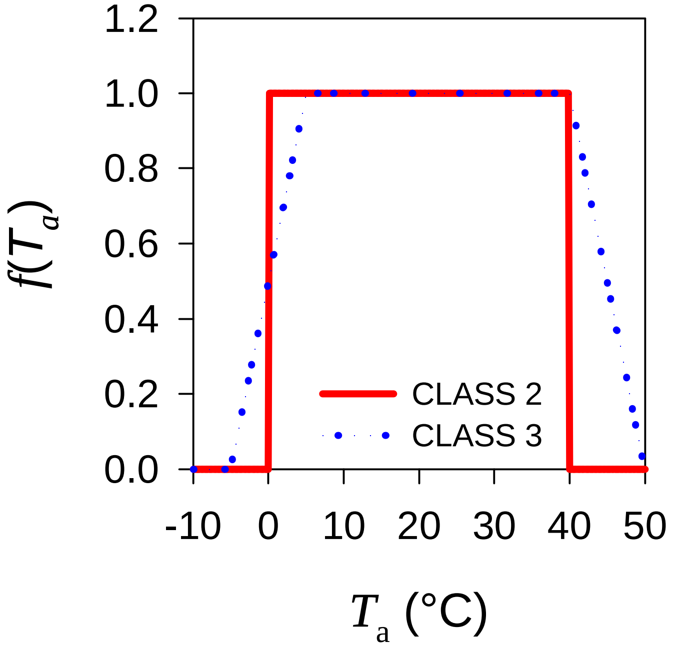
$$g_c = \frac{g_{s,\max}}{c_Q} \ln\left[\frac{Q_{\downarrow} + Q_{1/2}}{Q_{\downarrow}\exp\left(-c_Q\Lambda\right) + Q_{1/2}}\right] \cdot f\left(\Delta e\right) \cdot f\left(y_{s,r}\right) \cdot f\left(T_a\right)$$

- To represent various vegetation types, $f(\Delta e)$ and $f(\psi_{s,r})$ have adjustable coefficients that, along with $Q_{1/2}$, can be read from the initialization file, while default values are provided for major vegetation categories.

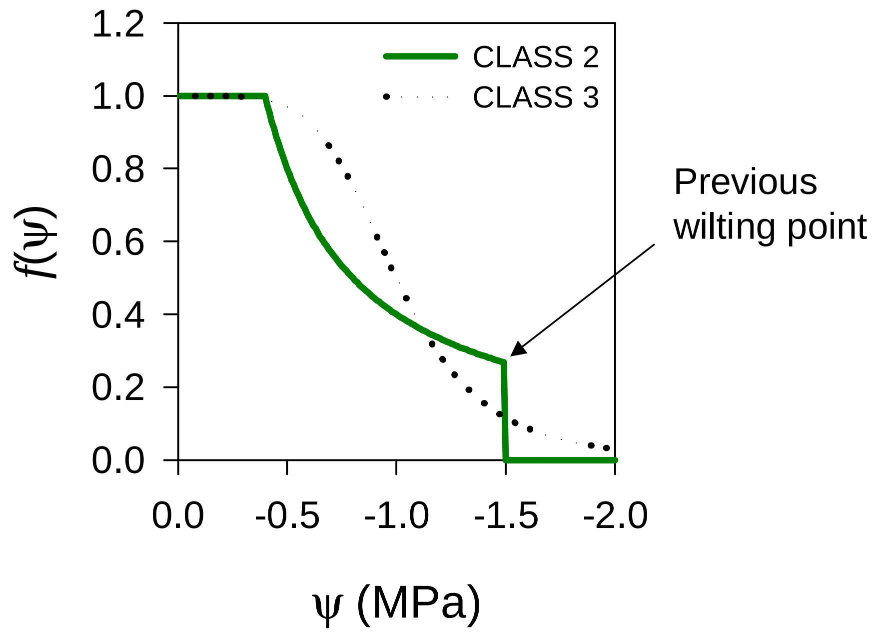
Canopy Conductance in CLASS 3.0

- To prevent step changes in CLASS's output:

- $f(T_a)$ has been changed, using more gradual bounds.



- The step change at a wilting point has been removed from $f(\psi_{s,r})$



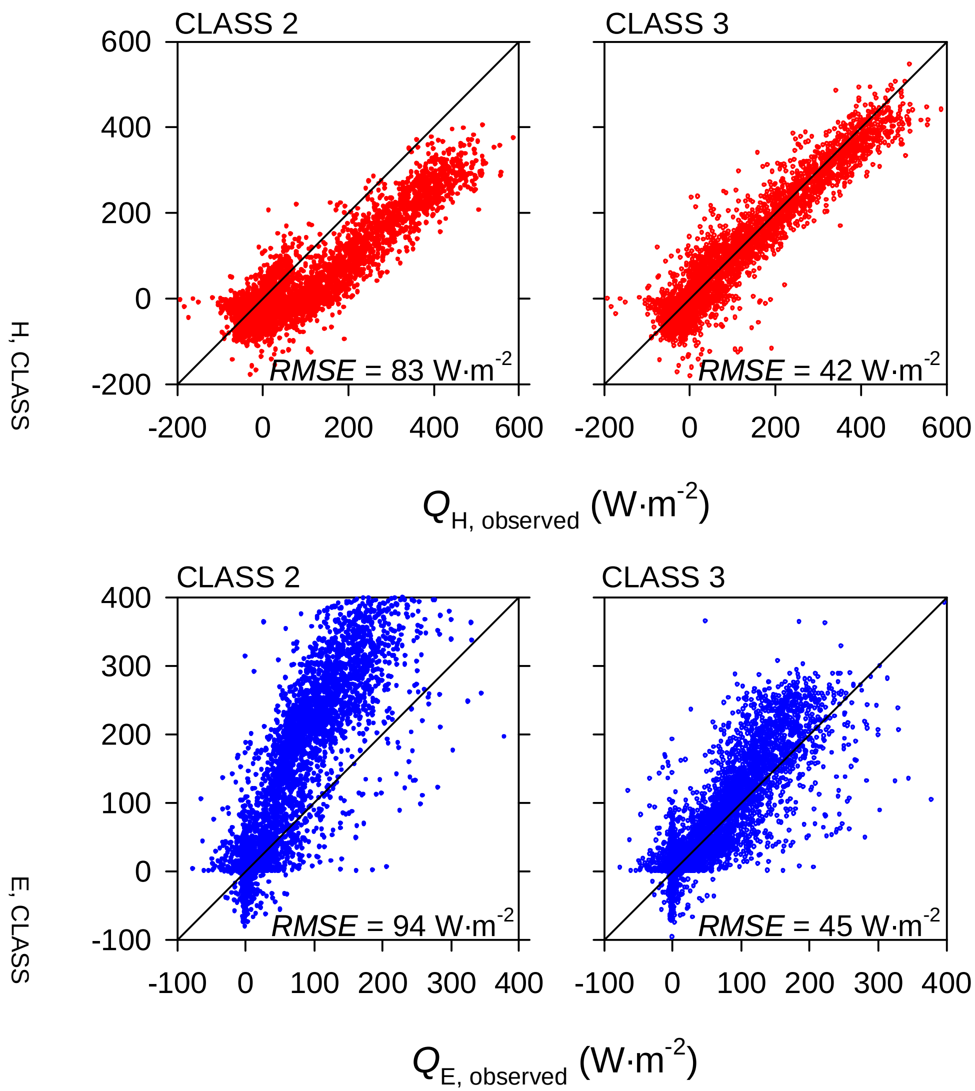
Performance of canopy conductance code for CLASS 2 and 3 at a boreal old black spruce stand in Manitoba

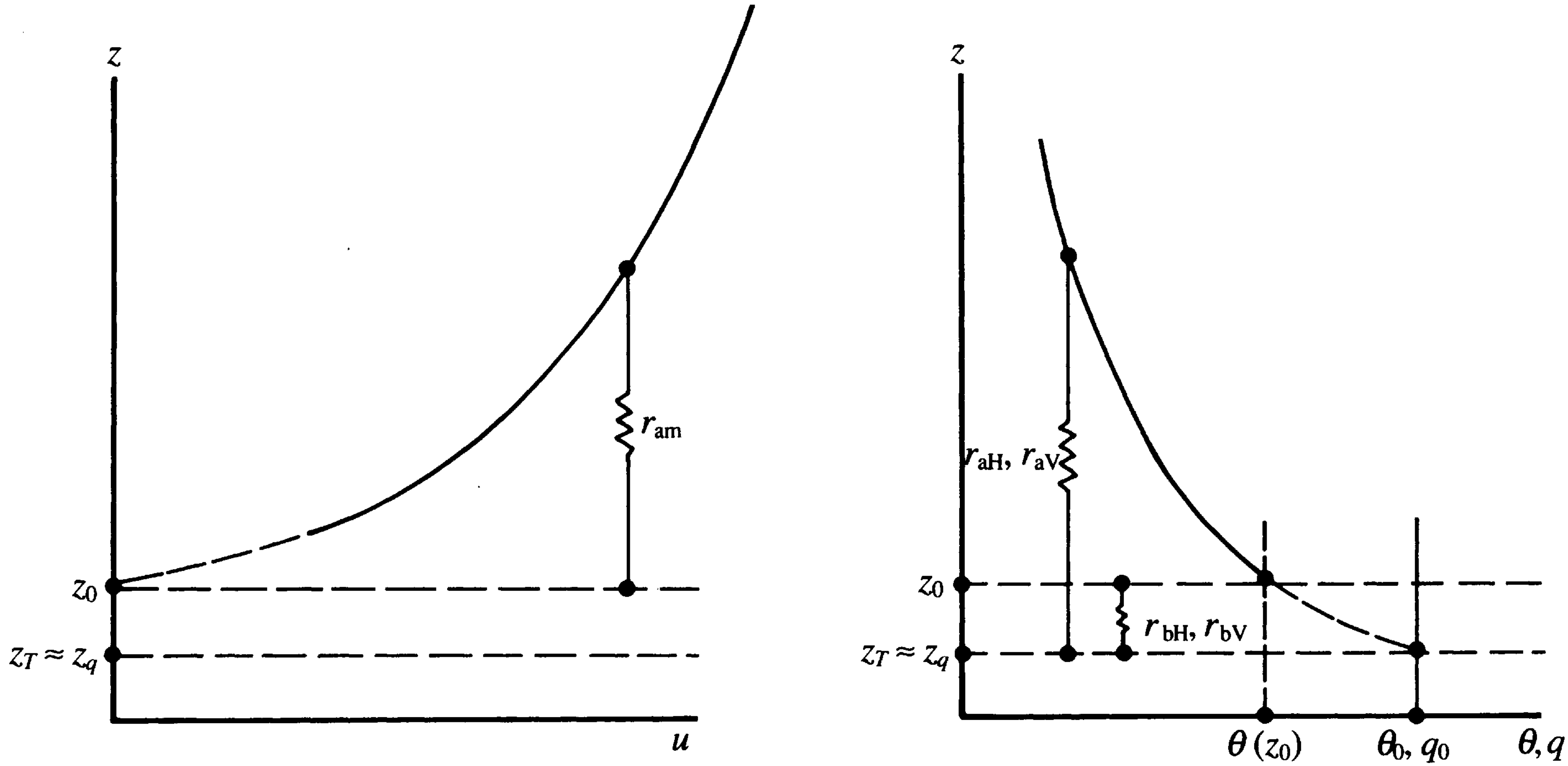
- In CLASS 2, canopy conductance was modelled using the same hard coded algorithm for all vegetation.

- Modelled canopy conductance was too large in the boreal forest.

- CLASS 2 overestimated the evaporation rate from boreal forests, and underestimated the sensible heat flux (left).

- In CLASS 3, the updated canopy conductance algorithm can represent a variety of vegetation types, and produces more realistic fluxes of heat and water (right).

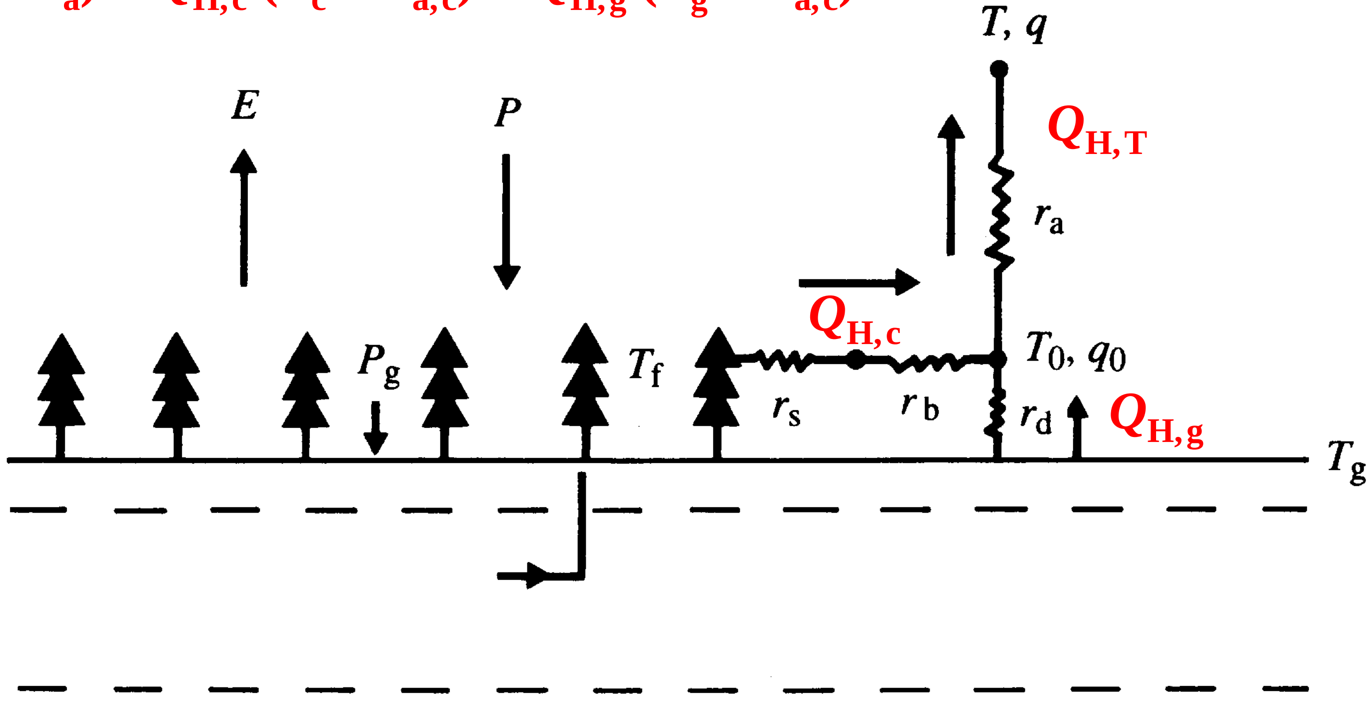




Schematic representation of aerodynamic resistances to the transfer of momentum and to the transfer of scalar properties, showing the excess resistance r_b due to molecular effects and the relation between the surface temperature θ_0 and the temperature $\theta(z_0)$.

(From J.R. Garratt, "The Atmospheric Boundary Layer")

$$Q_{H,T}\,(T_{a,c} - T_a) = Q_{H,c}\,(T_c - T_{a,c}) + Q_{H,g}\,(T_g - T_{a,c})$$



Schematic representation of the main elements of a non-isothermal or two-component canopy model. Linked to the atmosphere (via resistances r_s, r_b and r_a), to the soil or undergrowth (via resistance r_d) and the deep soil (via evapotranspiration), the canopy and upper soil layer are at temperatures T_f and T_g. P_g is the precipitation reaching the soil surface.

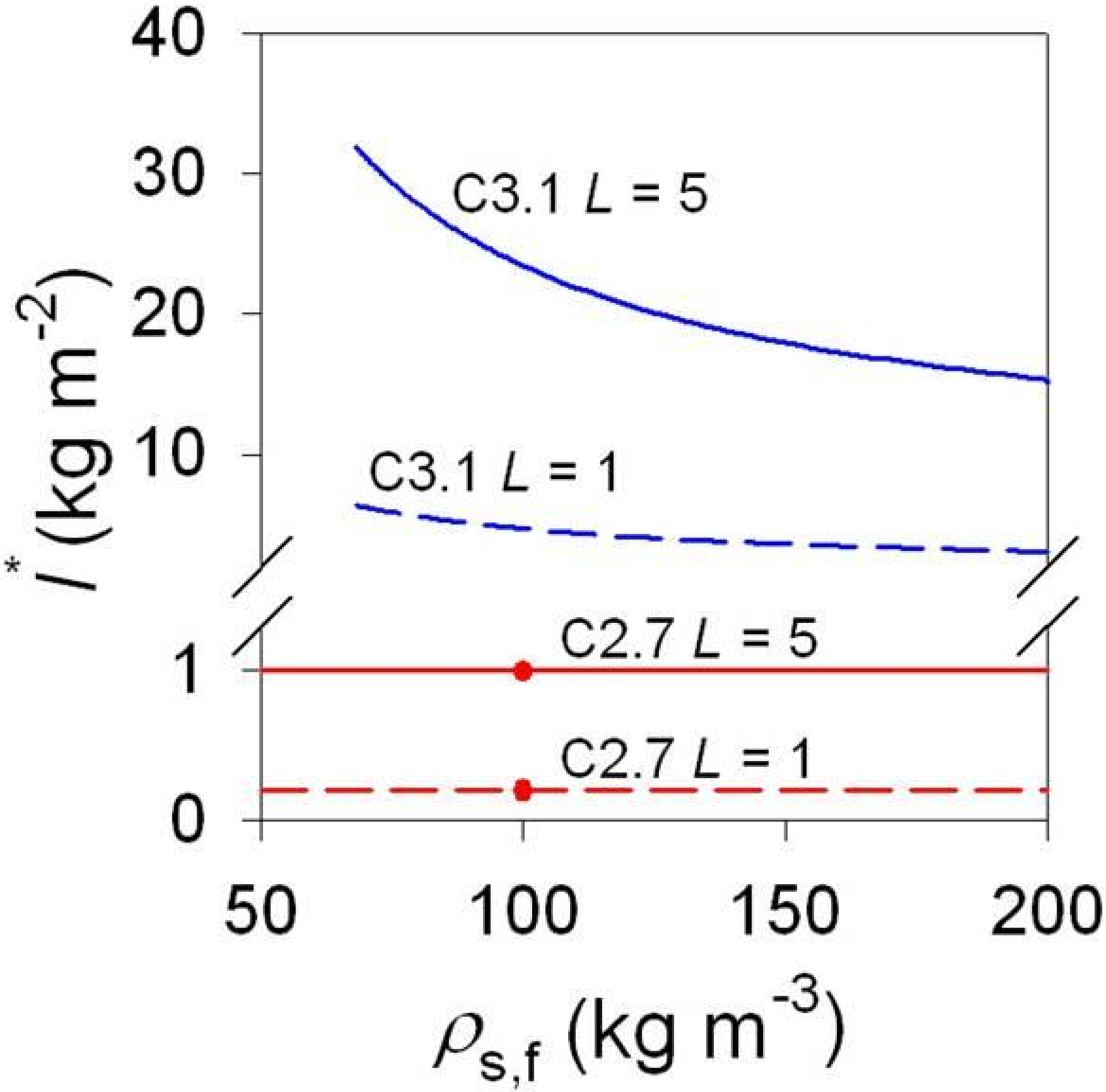
(From J.R. Garratt, "The Atmospheric Boundary Layer")

BERMS Old Black Spruce canopy (from tower)

Conifers have a much larger interception capacity for snow than for water

Following sublimation and unloading, the canopy is snow-free for much of the winter

Interception capacity

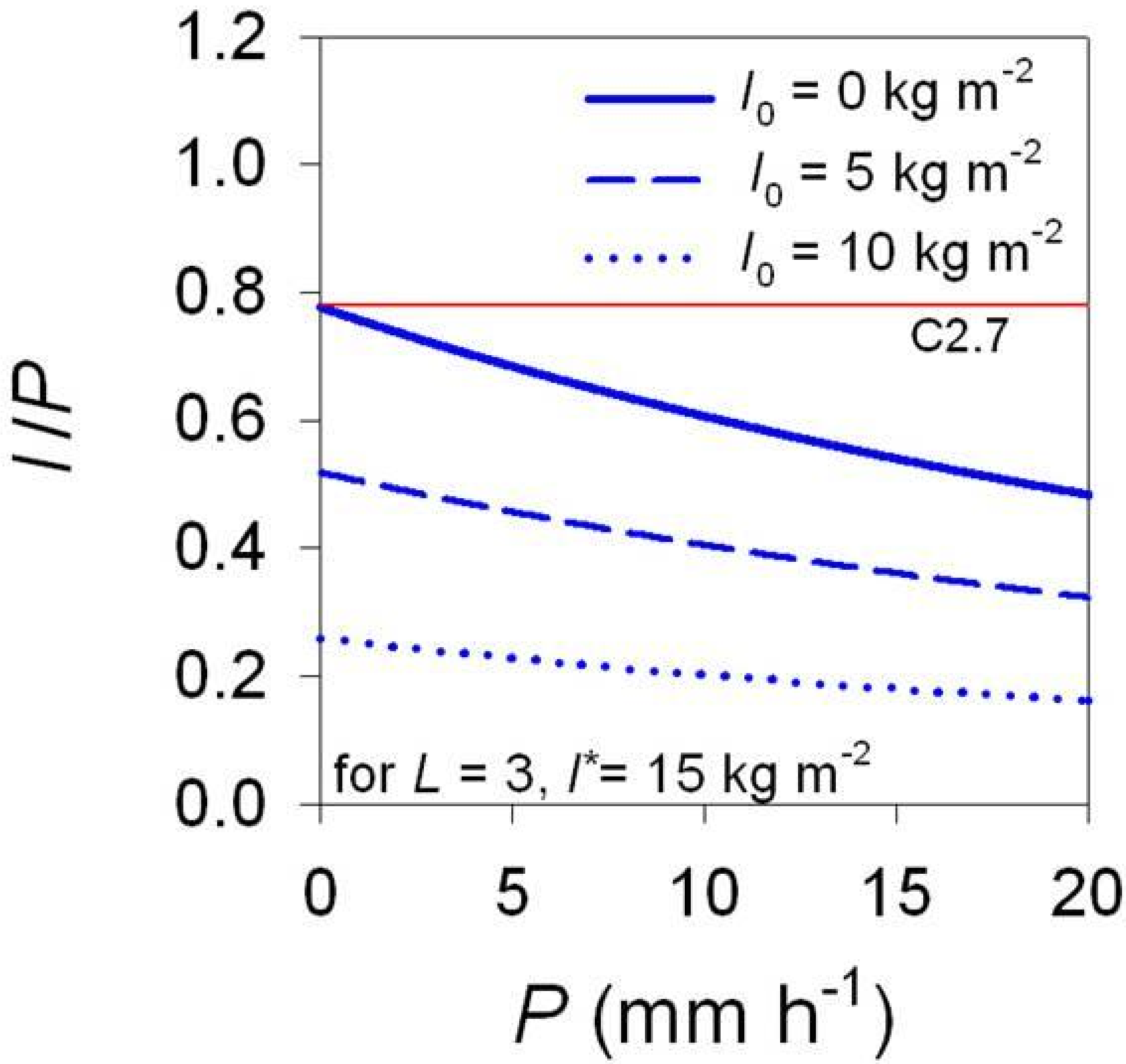


CLASS 2.7:
Interception capacity for snow, *I**, is treated the same as for rain.

CLASS 3.1
*I** is an order of magnitude larger for snow, but varies with snow density, $r_{\text{fresh snow}}$, in addition to projected leaf area index, *L*.

(Hedstrom and Pomeroy, 1998)

Interception efficiency

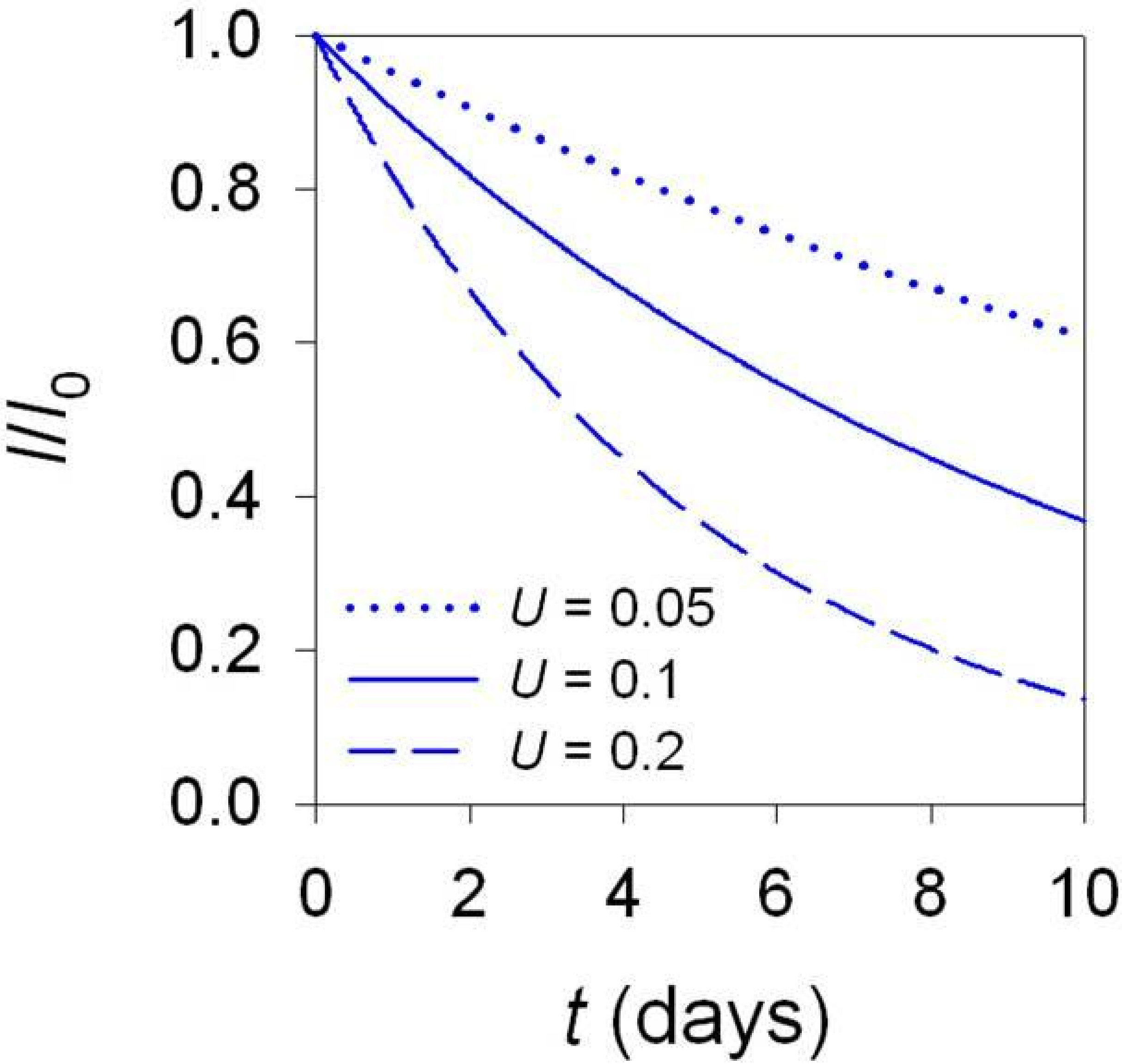


CLASS 2.7:
All snow landing on canopy is intercepted until the interception capacity, *I**, is reached.

CLASS 3.1:
Interception efficiency, *I*/*P*, decreases with precipitation rate, *P*, and also with the initial snow load on the canopy I_0.

(Hedstrom and Pomeroy, 1998)

Unloading of intercepted snow



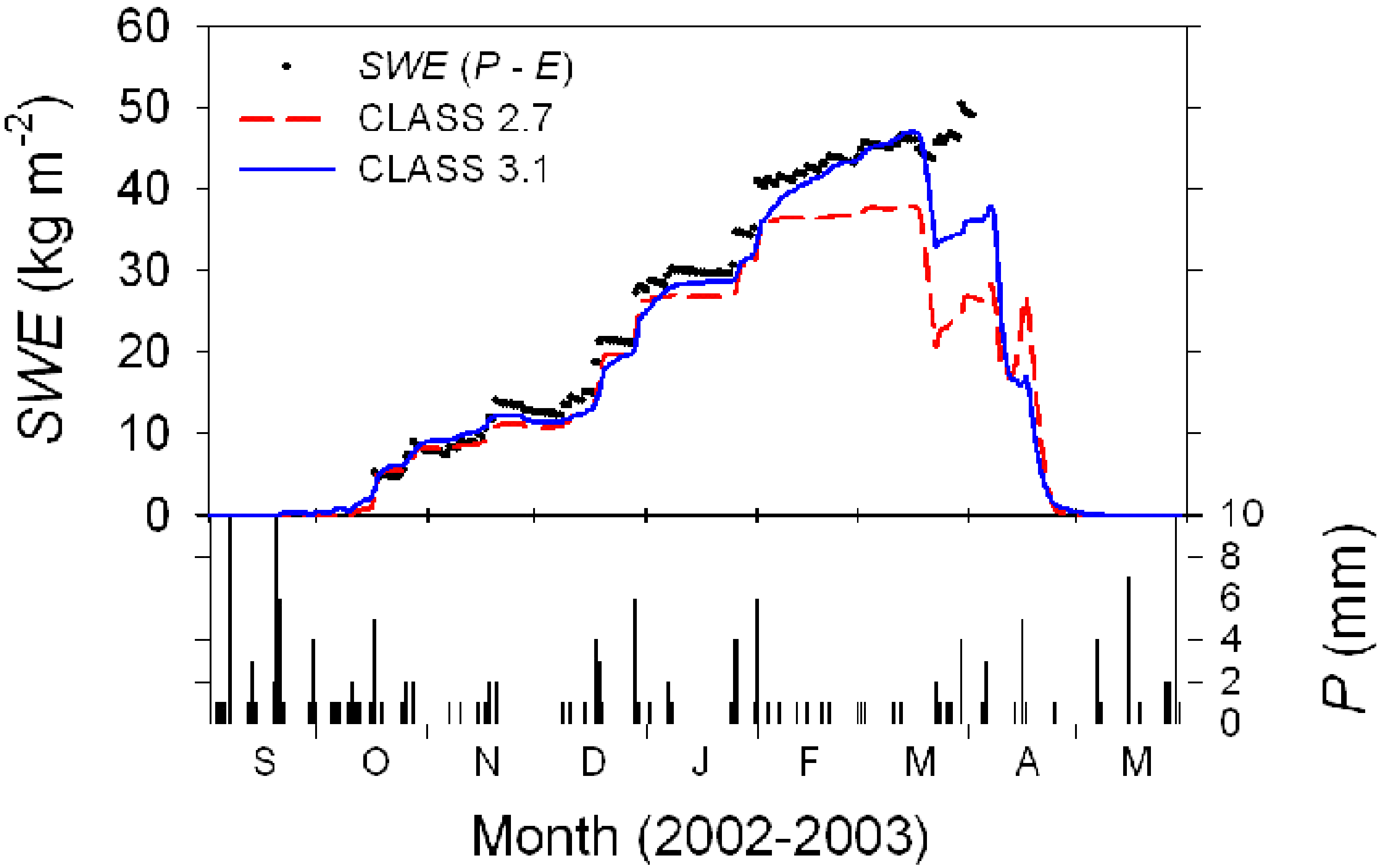
CLASS 2.7:
All intercepted snow remains on the canopy until it sublimates or melts.

CLASS 3.1:
Intercepted snow load, *I*, decreases over time according to the unloading rate coefficient *U* ($days^{-1}$).

Snow that unloads from the canopy is added to the snowpack.

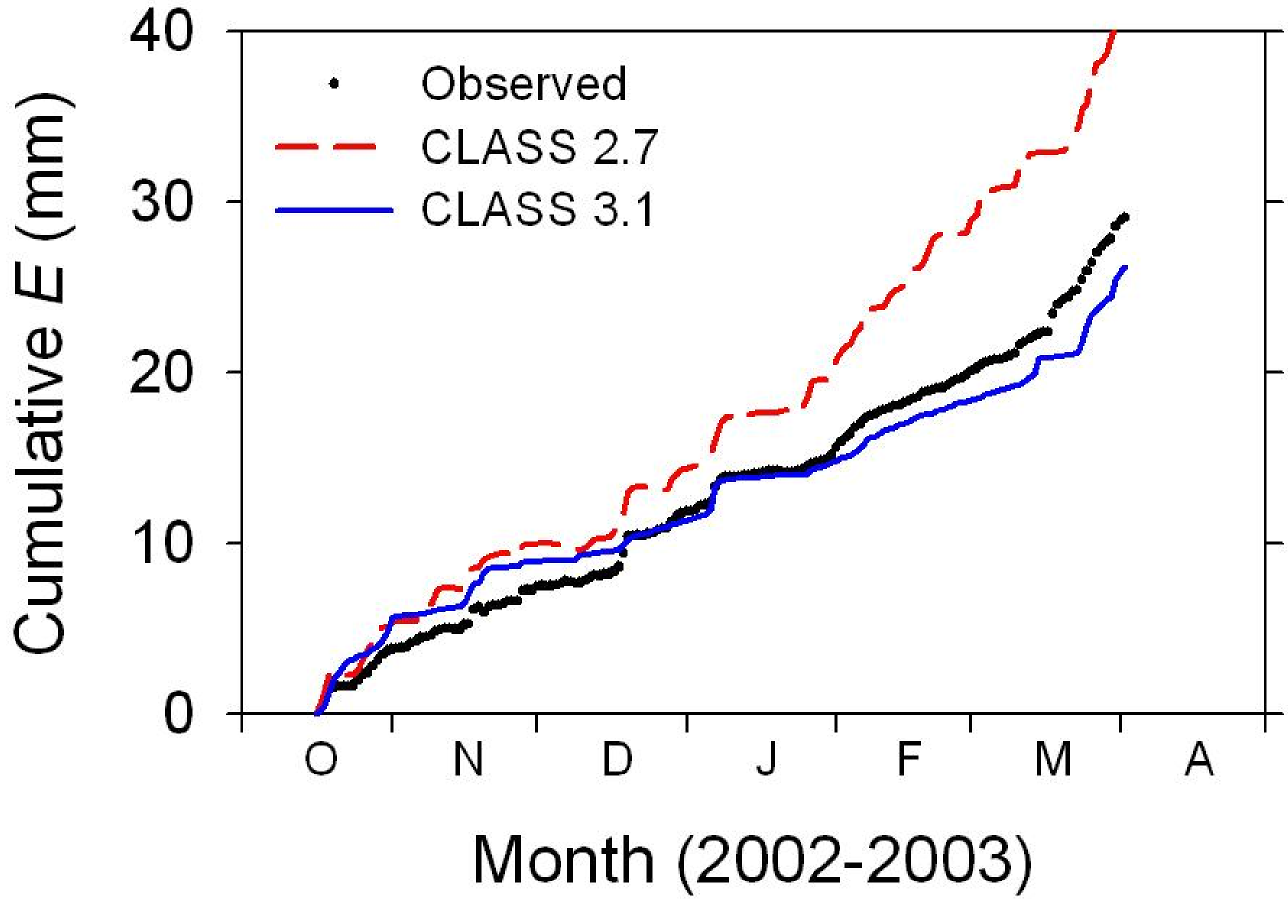
(Hedstrom and Pomeroy, 1998)

Observed and modelled snow water equivalent at the BERMS Old Black Spruce stand (2002-2003)



- CLASS 3.1 shows better agreement with measurements
- Most of the improvement in this model run comes from the ability to unload snow from small snowfall events before the snow sublimates.

Observed and modelled cumulative latent heat flux at the BERMS Old Black Spruce stand (2002-2003)



- CLASS 2.7 overestimates sublimation losses while CLASS 3.1 performs better.
- CLASS 3.1 is able to unload snow from the canopy to the forest floor where it is sheltered and less likely to sublimate.

The overall approach described above is the standard one used in atmospheric models to calculate the vapour flux to the atmosphere, and accounts for the effects of mechanical turbulence and buoyancy over a homogeneous or “well-mixed” surface in equilibrium with the overlying air.

Given that the horizontal grid resolution at the land surface that is used operationally in climate and weather forecast models is typically of the order of tens to hundreds of kilometers, the assumption that edge effects can be neglected in such applications is a reasonable one.

However, as grid resolutions increase, this assumption must increasingly be re-examined.

Effect of a discontinuity on atmospheric flow (from Monteith, 1973)

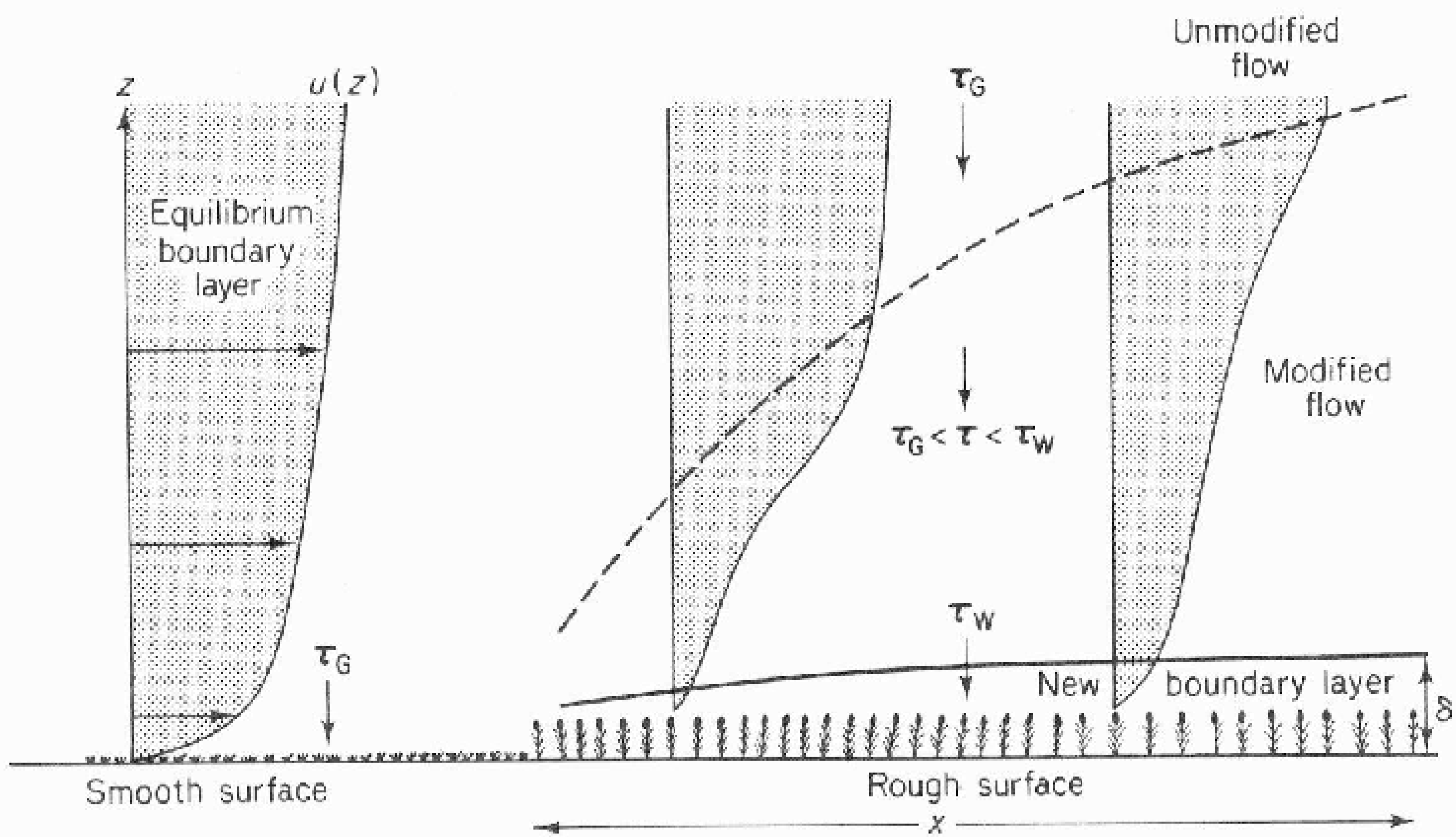
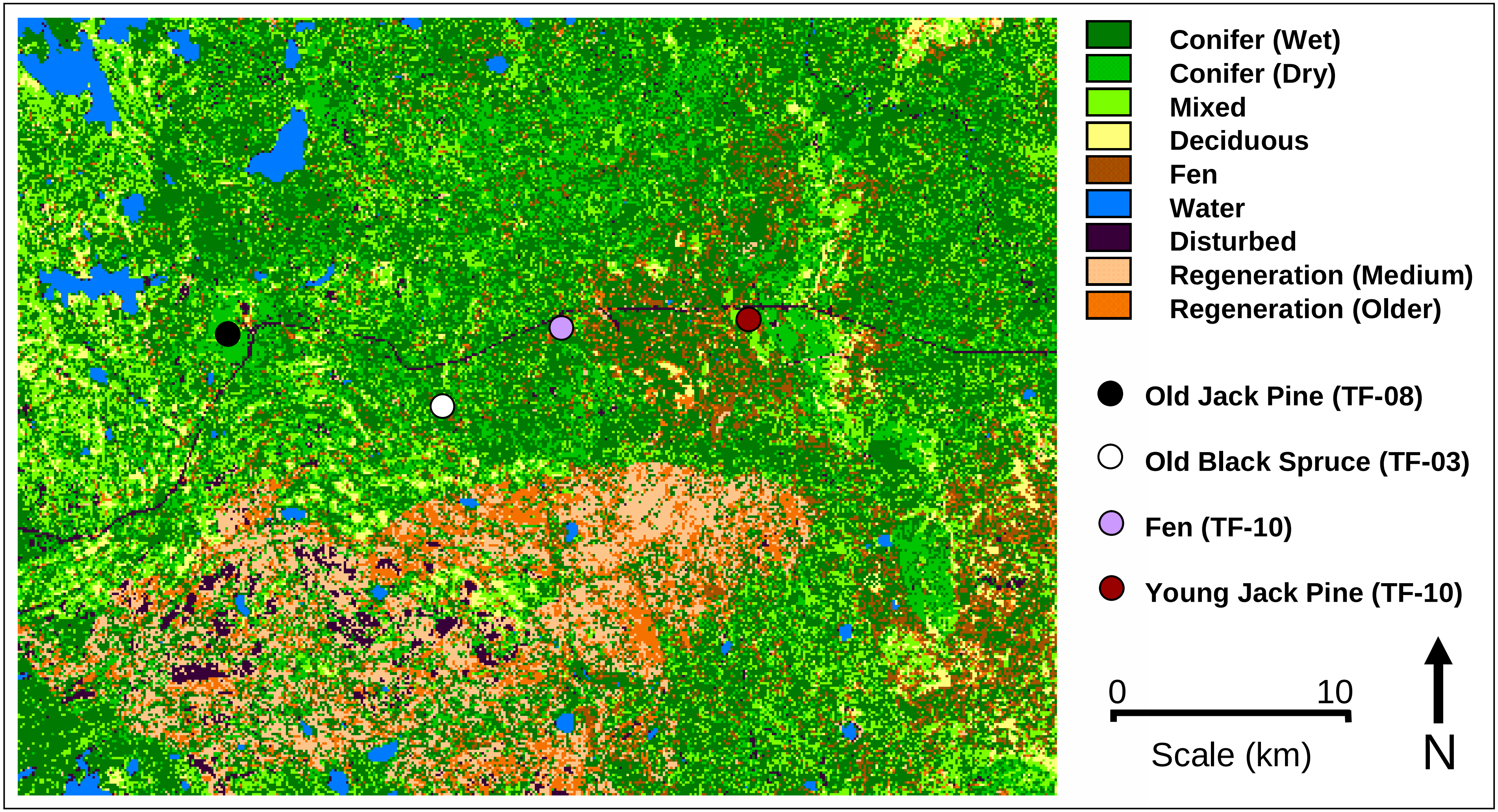
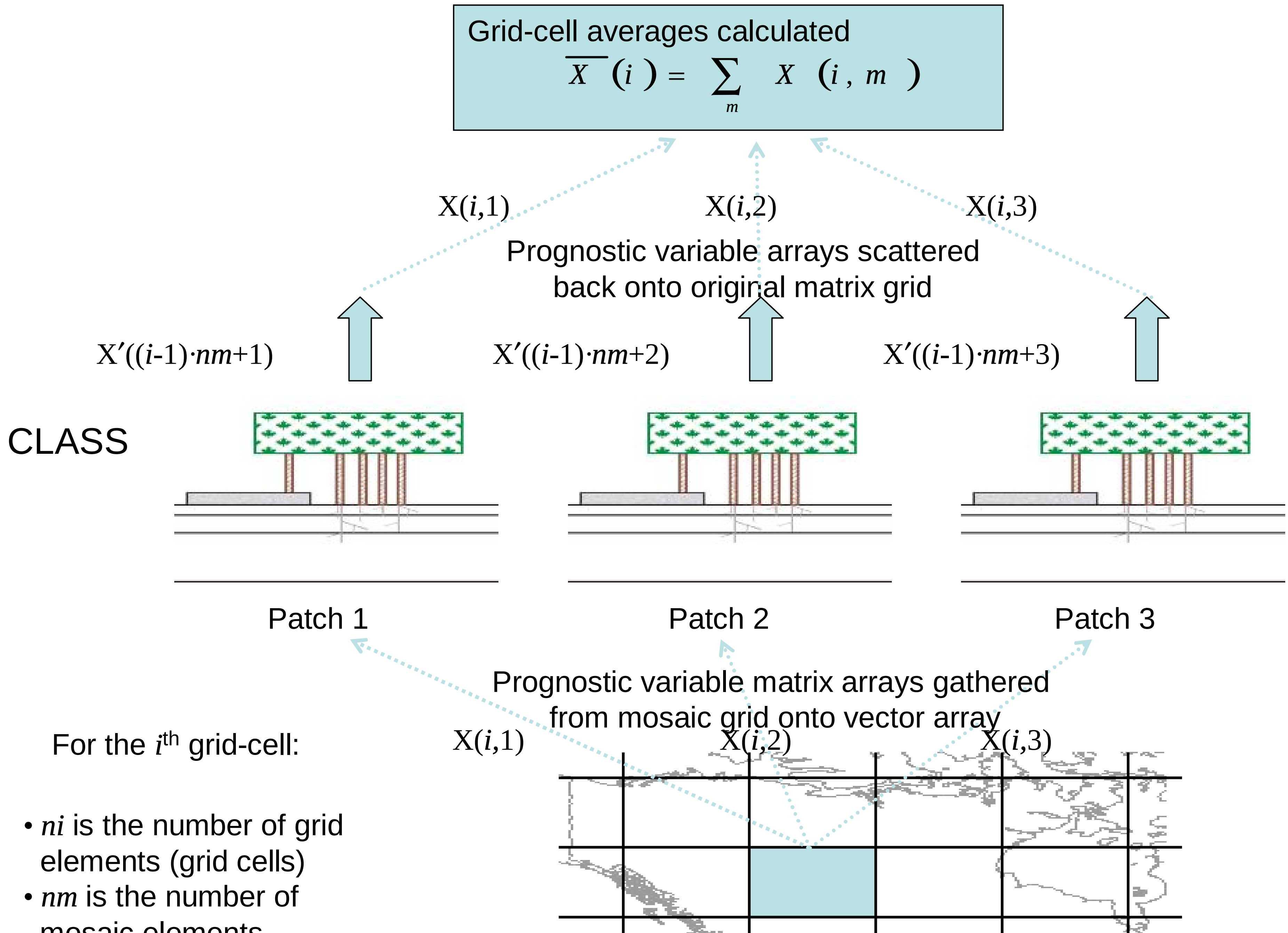


Fig. 6.9 Development of a new equilibrium boundary layer when air moves from a relatively smooth to a rougher surface. The ratio of the vertical to the horizontal scale is 20 :1. The broken line is the boundary between unmodified flow in which the vertical momentum flux is τ_G and modified flow in which the flux is between τ_G and τ_w. The flux is τ_w below the height δ.

Classification of study area from LANDSAT Thematic Mapper image



Forest cover classification by BOREAS group TE-18

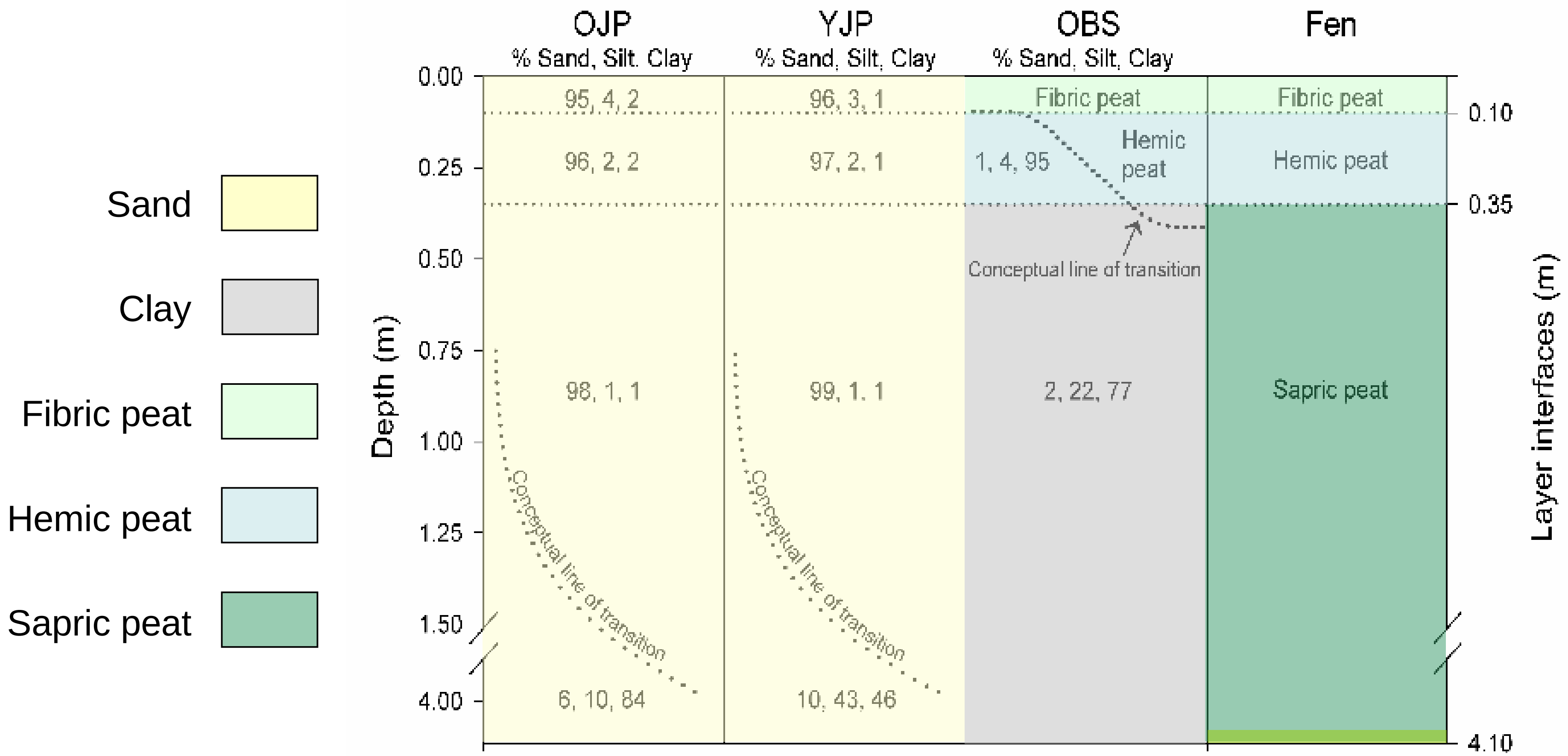


For the *i*th grid-cell:

- *ni* is the number of grid elements (grid cells)
- *nm* is the number of mosaic elements (patches in each grid-cell)

Site properties at each of the four flux tower sites

	OJP	YJP	OBS	FEN
Canopy height (m):	13.5	3	12	0.5
Leaf area index:	1.9	1.2	4.6	1.5
RCMIN ($s \cdot m^{-1}$):	165	200	270	130

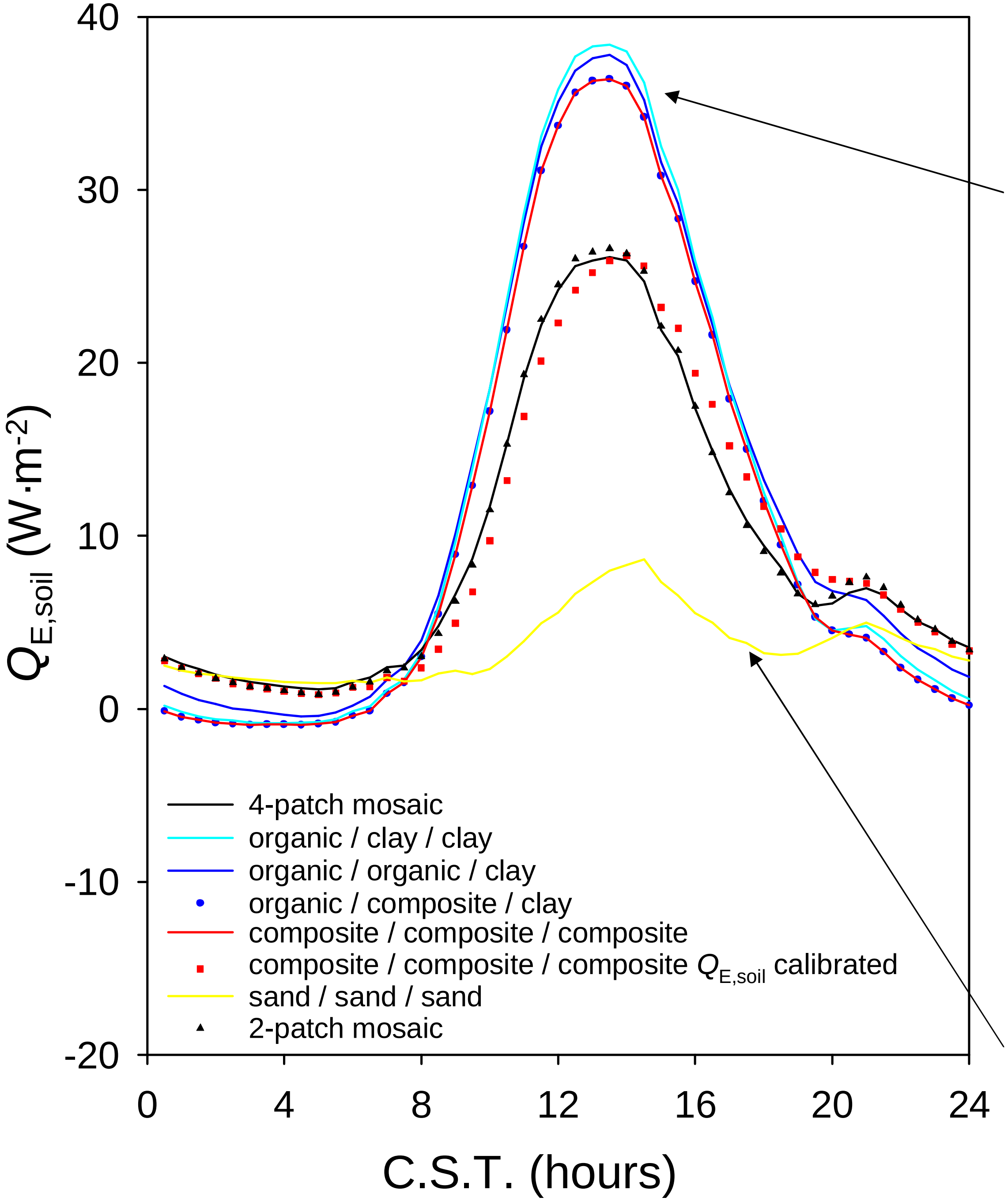


Soil columns employed in the aggregated and mosaic runs

	Layer 1	Layer 2	Layer 3	Comment
1.	Organic	Clay	Clay	exists in region
2.	Organic	Organic	Clay	OBS
3.	Organic	Composite	Clay	
4.	Composite	Composite	Composite	Weighted average of four tower-flux sites.
5.	Composite	Composite	Composite	$Q_{E,soil}$ calibrated
6.	Sand	Sand	Sand	YJP and OJP

2-patch mosaic:
1. Sand patch (3 sand layers)
2. Organic/clay patch (2 organic layers over a clay layer)

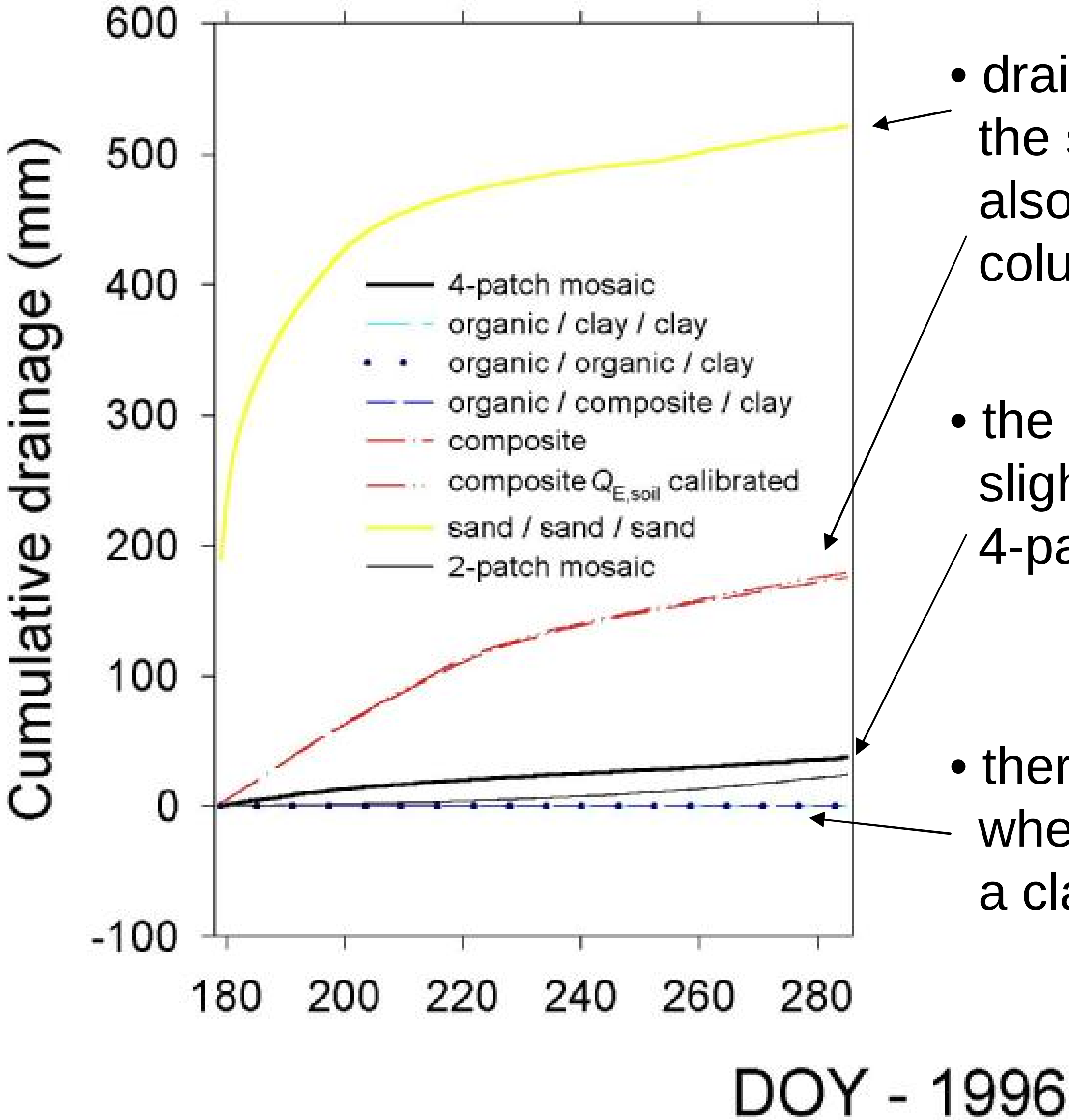
4-patch mosaic:
- 3 sand layers (YJP)
- 3 sand layers (OJP)
- 2 organic layers over a clay layer (OBS)
- 3 organic layers (Fen)



Soil evaporation

- With an organic soil layer or a composite soil layer at the surface, $Q_{E,soil}$ is overestimated.
- The 4-patch and 2-patch mosaic model runs have almost identical behaviour.
- We were able to tune the algorithm for soil evaporation so that the composite soil behaved similarly to the mosaic.
- This is not meant as a solution to soil evaporation, but merely to minimize one source of error while examining modelled Q_E.
- With a sand soil column, $Q_{E,soil}$ is underestimated due to rapid drainage.

Cumulative drainage and Q_E for aggregated model runs employing various soil columns, and for 4- and 2-patch mosaic model runs



- drainage is much too rapid in the sand soil column, and is also rapid in the composite soil column
- the 2-patch mosaic drains slightly more slowly than the 4-patch mosaic
- there is very little drainage when the third soil layer is a clay