Field Observation in Climatology and Environmental Hydrology

Module: Energy and mass exchange between the Earth's surface and the atmosphere

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Learning objectives

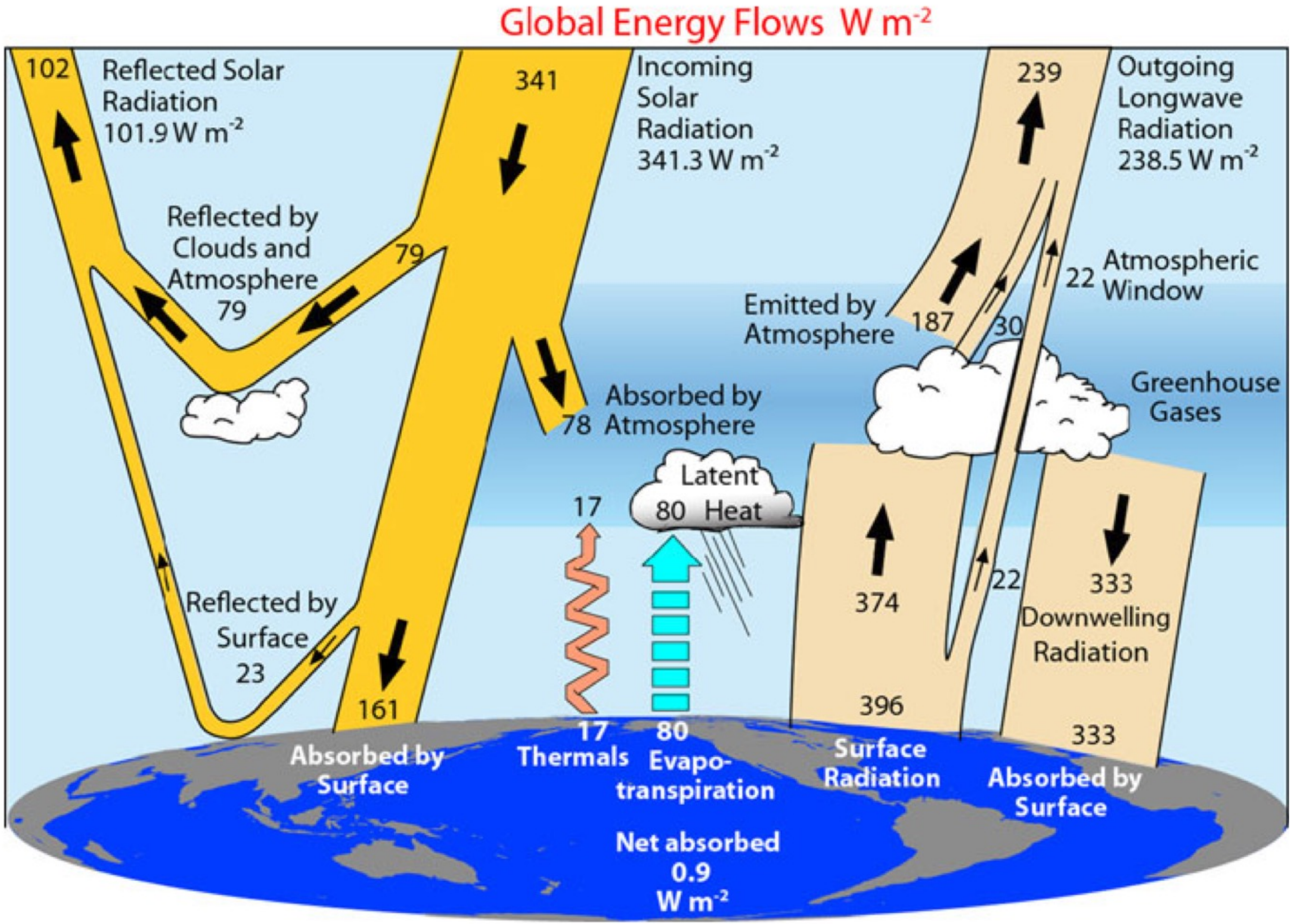
- Properties of the Earth’s surface
- Energy and mass fluxes at the surface
- Energy balance terms
- Parametrizations of fluxes

Why bother with the bottom boundary?

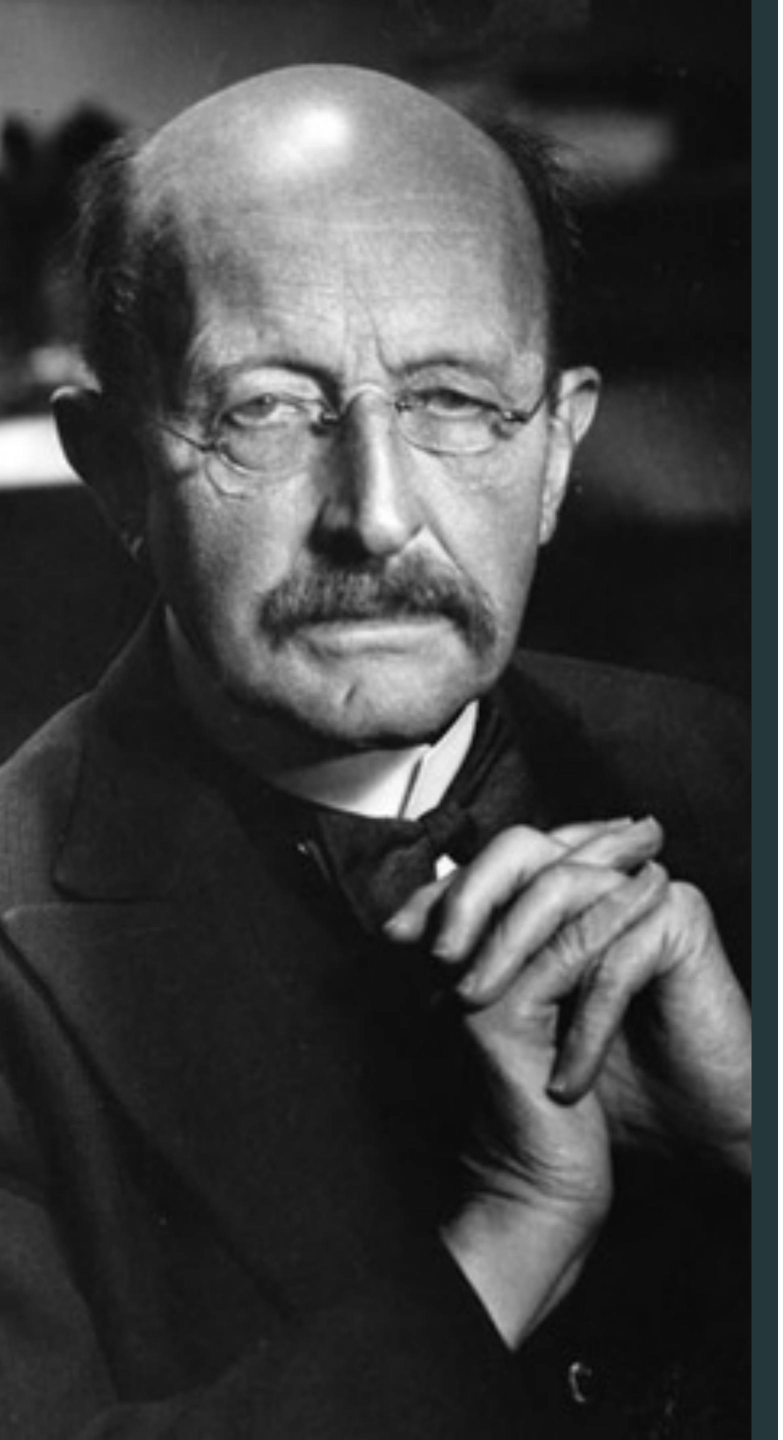
… because

- without a bottom boundary there would be no boundary layer
- Friction, heat and moisture fluxes from the surface modify the state of the BL
- Heat and moisture fluxes are driven by the external forcings (radiation, transporation etc.)

The global energy budget



Trenberth and Fasullo (2012) Surv. Geophys.



PLANCK'S LAW OF RADIATION

<u>Planck (1900):</u>

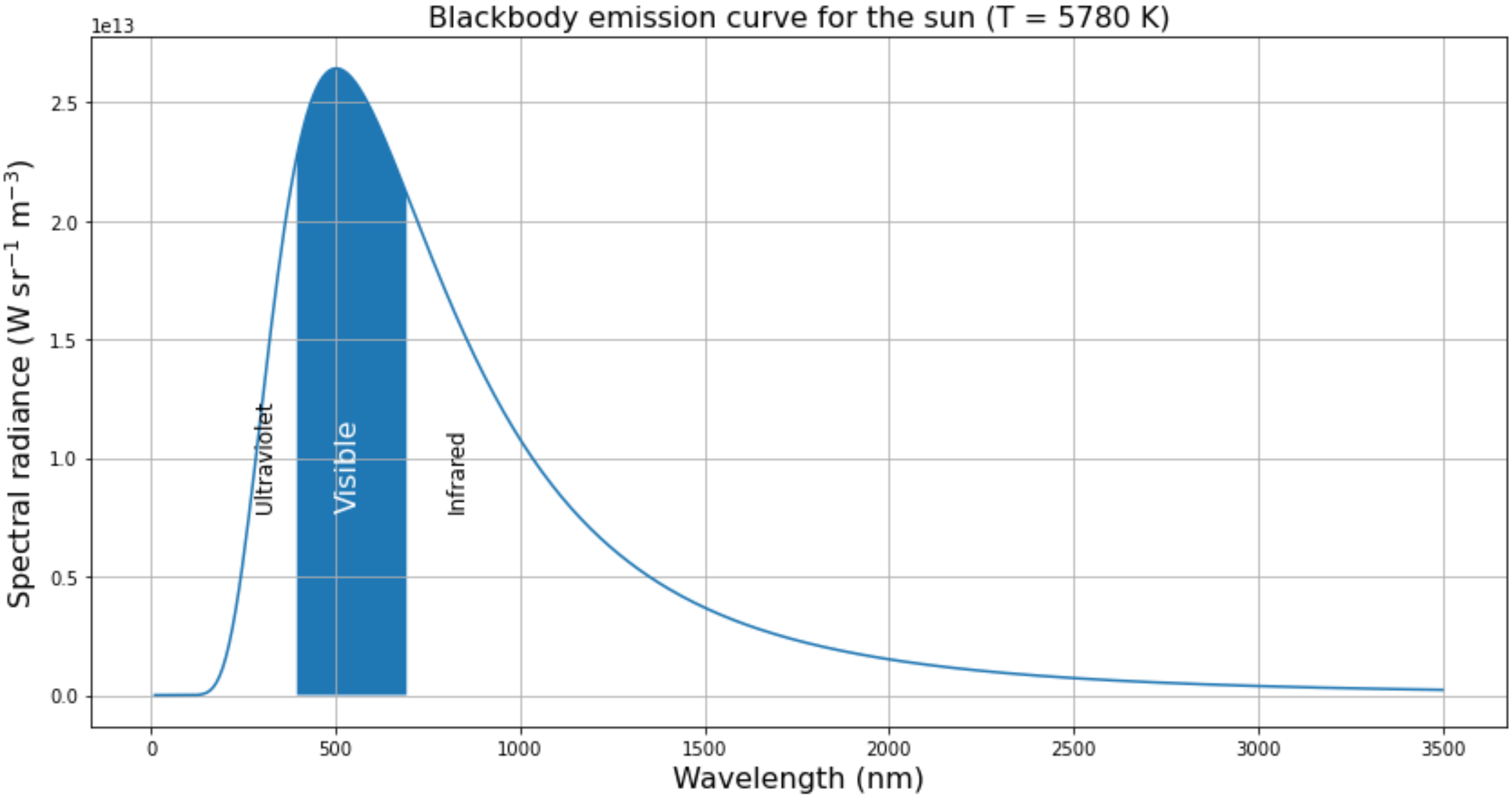
Every body emits thermal radiation (energy), which depends on the temperature of the radiating body.

<u>Blackbody radiator:</u>

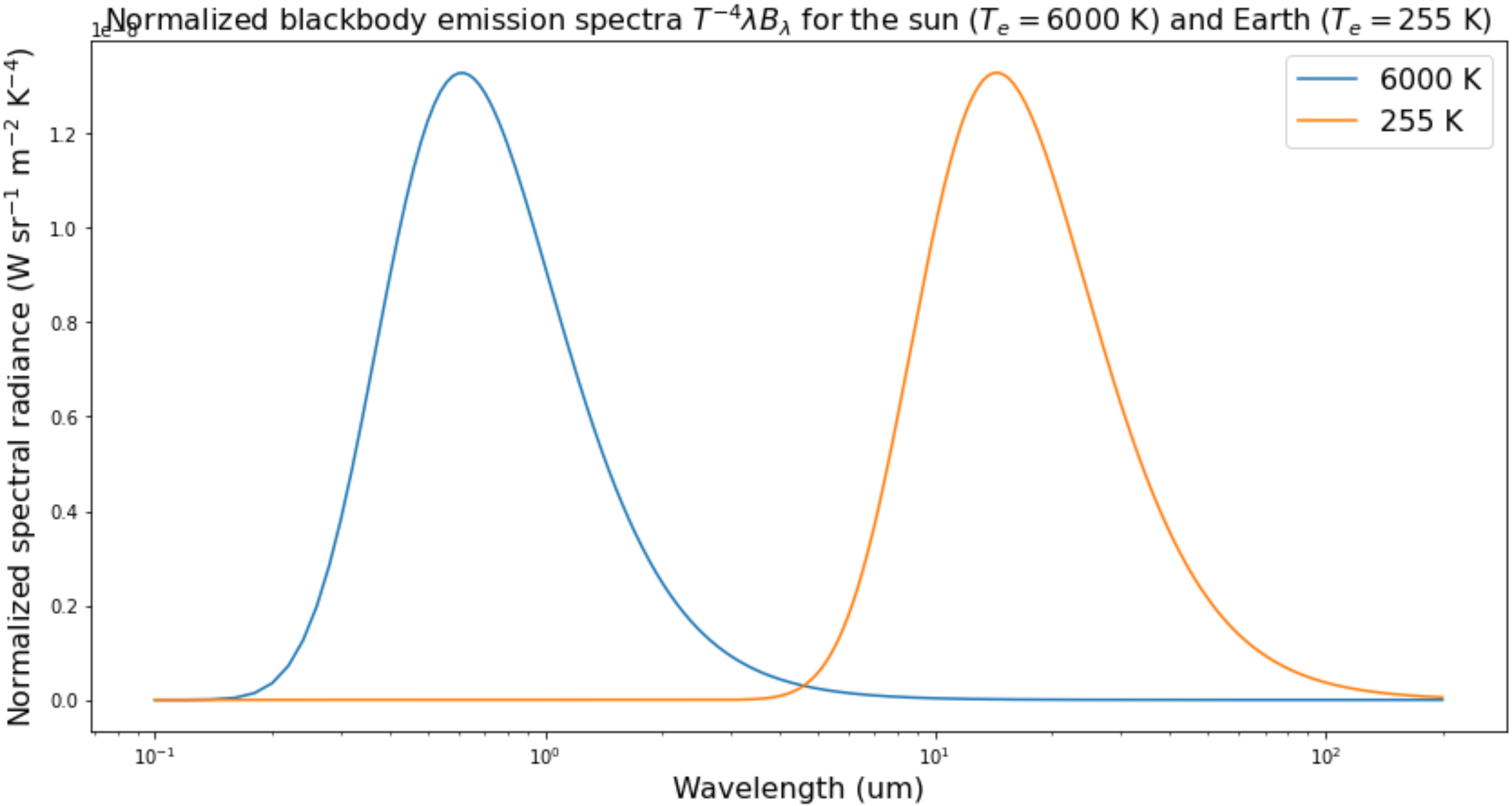
- Body that completely absorbs incident electromagnetic radiation.
- It emits at the respective temperature with the maximum possible radiation intensity.
- In the mid- and far-IR range, natural land surfaces behave approximately like black bodies.

$$\frac{\partial Q}{\partial \lambda} = \frac{2hc^2}{\lambda^5 \left[\exp\left(\frac{hc}{k\lambda T}\right) - 1\right]}$$

$h \approx 6.626*10^{-34}$ Js (Plank'sche constant)
$k \approx 1.381*10^{-23}$ JK^{-1} (Boltzmann-Constant)
$c \approx 2.9979*10^{8}$ ms^{-1} (Speed of light in a vacuum)
λ = Wellenlänge

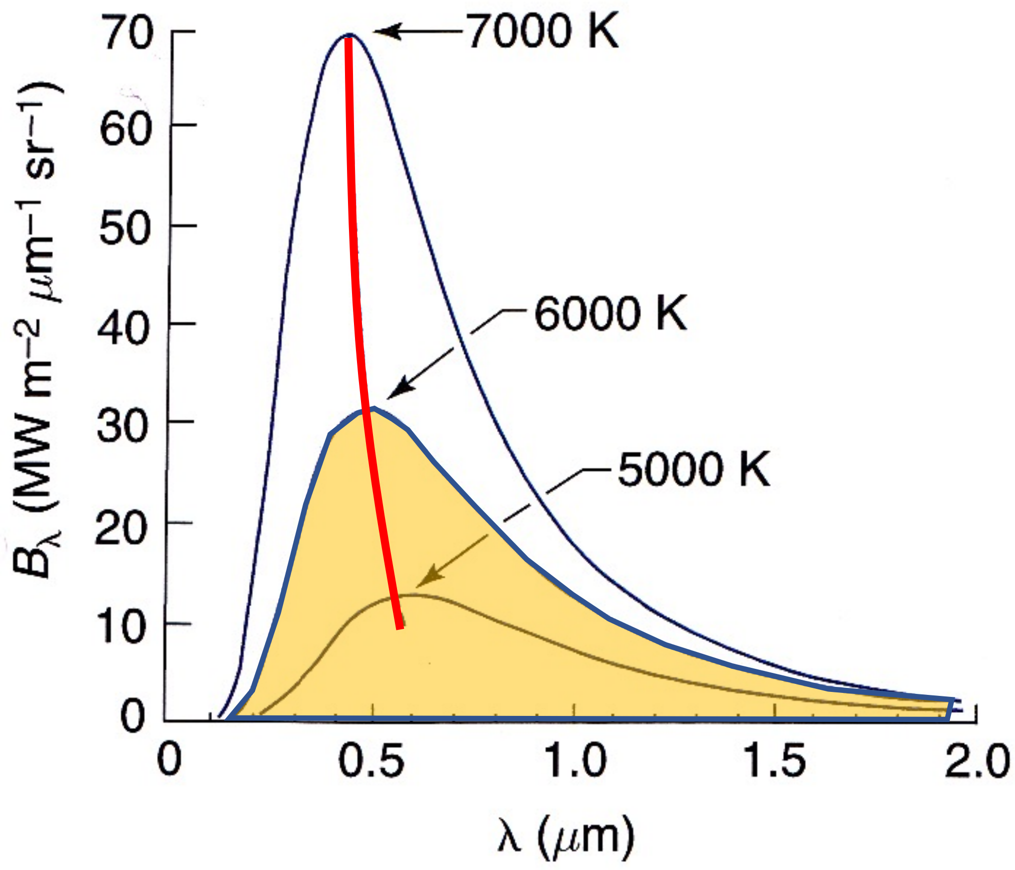


Source: Rose, 2020.



Source: Rose, 2020.

Stefan-Boltzmann Law



- The curves describe the distribution of the energy over the different wavelengths
- The areas under the curves describe the amount of radiated energy
- The radiated energy is temperature-dependent, namely to the 4th power of the absolute temperature, more precisely:

$$E = \sigma \cdot T^4 \qquad \text{mit } \sigma = 5.67 \cdot 10^{-8}\ [W/m^2K^4]$$

- → **Stefan-Boltzmann-Law**

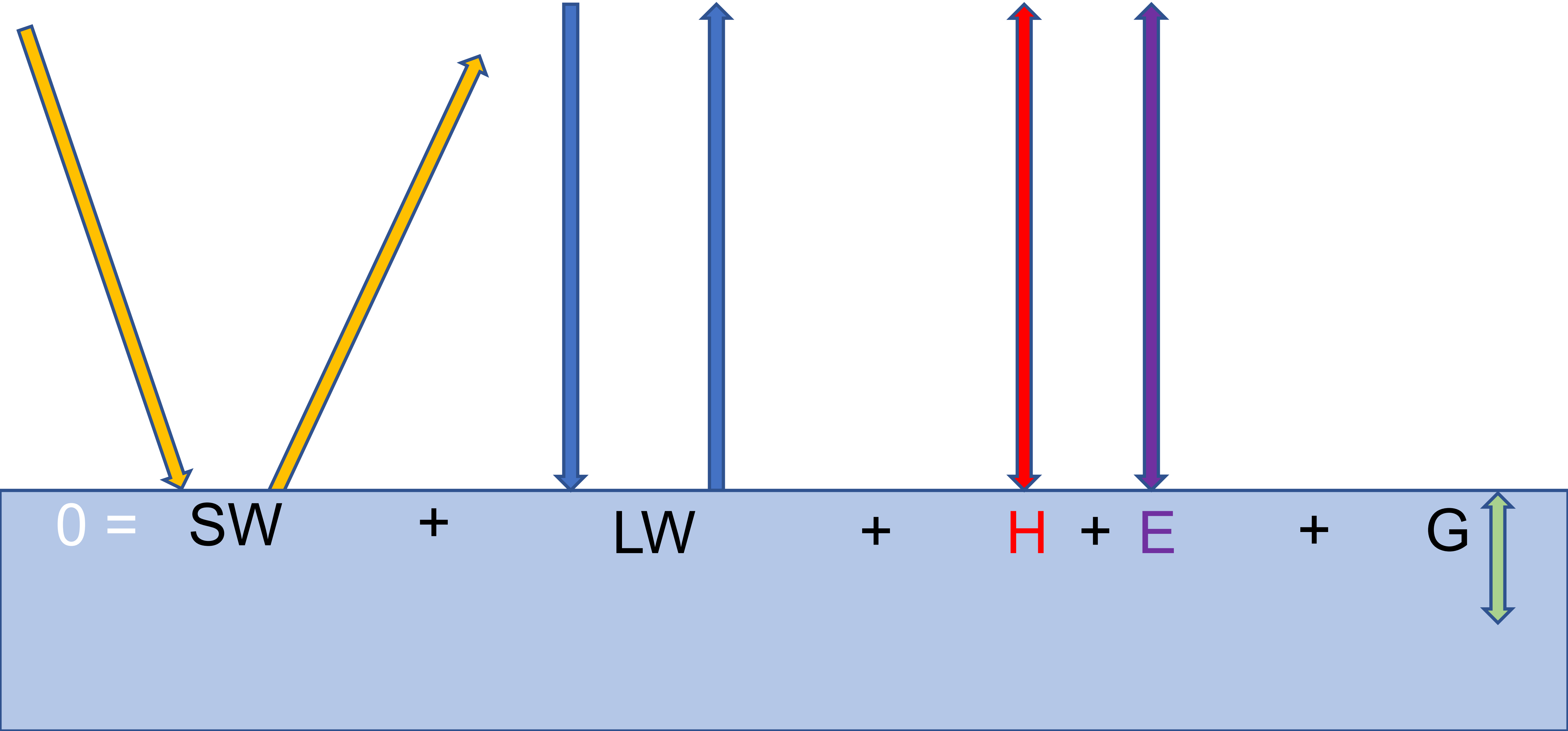
- The position of the emission maximum shifts to shorter and shorter wavelengths with increasing temperatures. The product of temperature and wavelength of the emission maximum is constant

$$\lambda_{max} \cdot T = 2880\ \mu m \cdot K$$

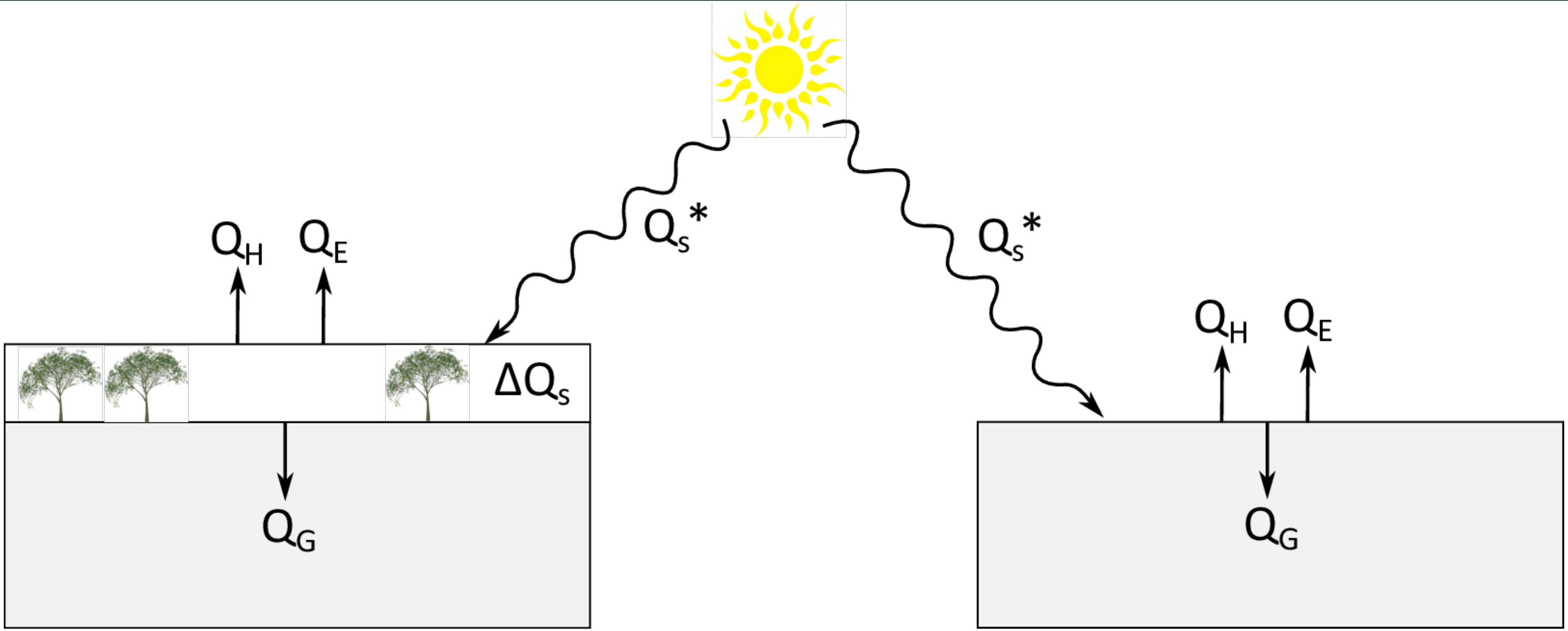
→ **Wien‘s law**

Quelle: Wallace & Hobbs 2006:117

Surface Energy Balance model (SEB)



Surface Energy Balance



$$Q_S^* = Q_H + Q_E + Q_G + \Delta Q_s$$

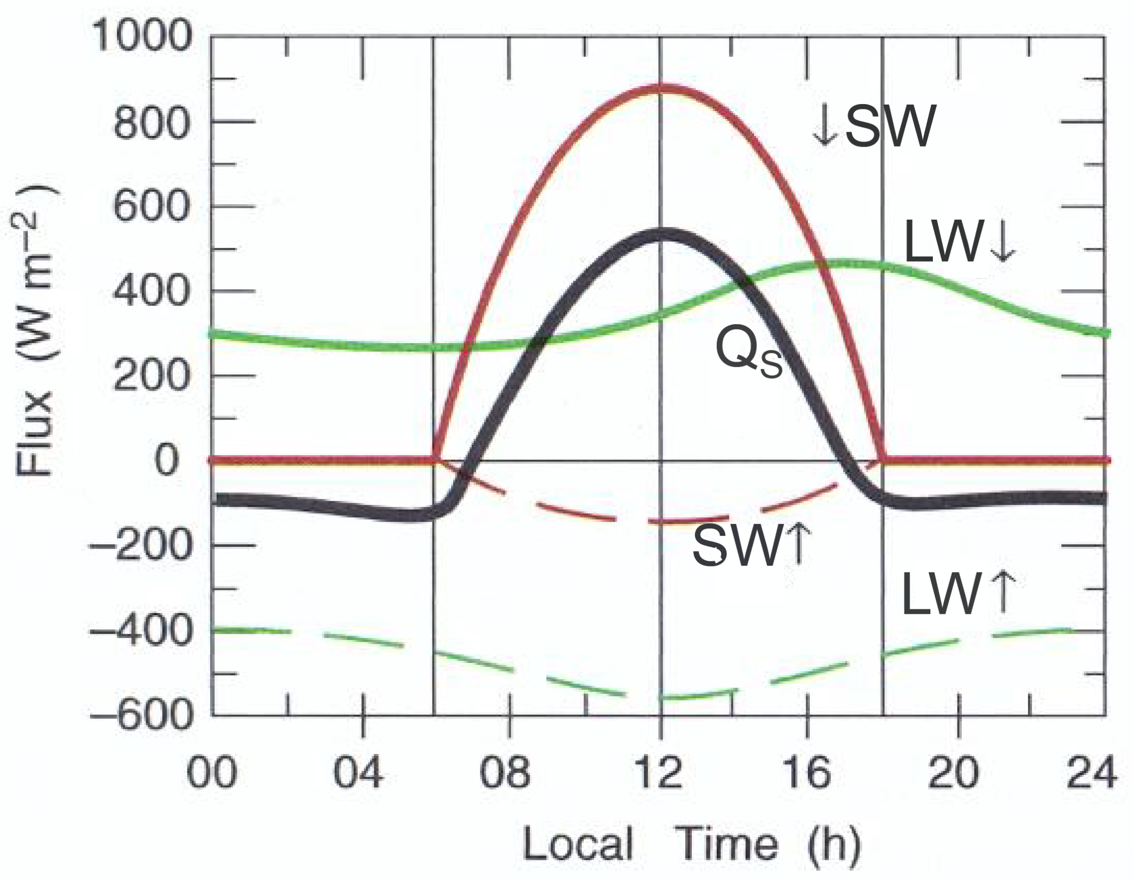
with Q_S^* the net radiation at the surface, Q_H the sensible heat flux, Q_E the latent heat flux, Q_G the ground heat flux, and ΔQ_s the storage term (internal energy)

FLUXES AWAY FROM SURFACE ARE POSITIVE, EXCEPT FOR NET RADIATION

Parametrization of radiation fluxes

$$Q_S^* = Q_H + Q_E + Q_G + \Delta Q_s$$

Parametrization of radiation fluxes



$$Q_S^* = SW_u + SW_d + LW_u + LW_d$$

with Sw_u : upwelling reflected shortwave radiation
SW_d: downwelling shortwave radiation transmitted through the air
LW_u: longwave (infrared, IR) radiation emitted up
LW_d: longwave diffusive IR radiation down

Shortwave radiation

$$\boldsymbol{SW_d = S_0 \cdot T_k \cdot \sin(\psi)}$$ for daytime (i.e. $\sin(\psi)$ is positive)

$$\boldsymbol{SW_d = 0}$$ for nighttime

with the transmissivity (parametrization by Burridge and Gadd, 1974)

$$T_k = (0.6 + 0.2 \sin\psi) \cdot (1 - 0.4\sigma_{C_H}) \cdot (1 - 0.7\sigma_{C_M}) \cdot (1 - 0.4\sigma_{C_L})$$

σ_C represents the cloud cover fraction, and H, M, and L signify high, middle, and low clouds

and the local elevation angle (Zhang and Anthes, 1982)

$$\sin\psi = \sin\phi \sin\delta_s - \cos\phi \cos\delta_s \cos\left[\left(\frac{\pi t_{UTC}}{12} - \lambda_e\right)\right]$$

ϕ : latitude (positive north) in radians
λ : longitude (positive west) in radians
t_{UTC} : Coordinated Universal Time in hours

δ_s is the solar declination angle (angle of the sun above the equator, in radians)

$$\delta_s = \phi_r \cos\left[\frac{2\pi(d - d_r)}{d_y}\right]$$

ϕ_r : latitude of the Tropic of Cancer (0.409 radians)
d : number of the day of the year
d_r : day of the summer solstice (173)
d_y : average number of days per year

Albedo

$$SW_u = \alpha \cdot SW_d$$

Surface	a in %
Fresh snow	75-95
Wet snow	60-70
Firn	40-70
Glacier ice	30-45
Impure Glacier ice	20-30
Ocean, lakes	6-12
Wet sand	15-30
Dry sand	25-40
Rocks	10-40
Concrete	10-35
Dark soil	5-10
Forest	10-20
Meadows and fields	10-30

Longwave radiation

$$\boxed{\boldsymbol{LW_d = \epsilon_a \cdot \sigma \cdot T_a^4}}$$

$$\boxed{\boldsymbol{LW_u = \epsilon_s \cdot \sigma \cdot T_0^4}}$$

with emissivities

$$\epsilon_a = 0.23 \cdot 0.433 \cdot \left(\frac{100 \cdot f \cdot E_{sat}(T_a)}{T_a}\right)^{1/8}$$

$$\epsilon_s = 1.0$$

with f the relative humidity, E_{sat} the saturation water vapour pressure, and T_a the air temperature.

Parametrization of ground heat flux

$$Q_S^* = Q_H + Q_E + Q_G + \Delta Q_s$$

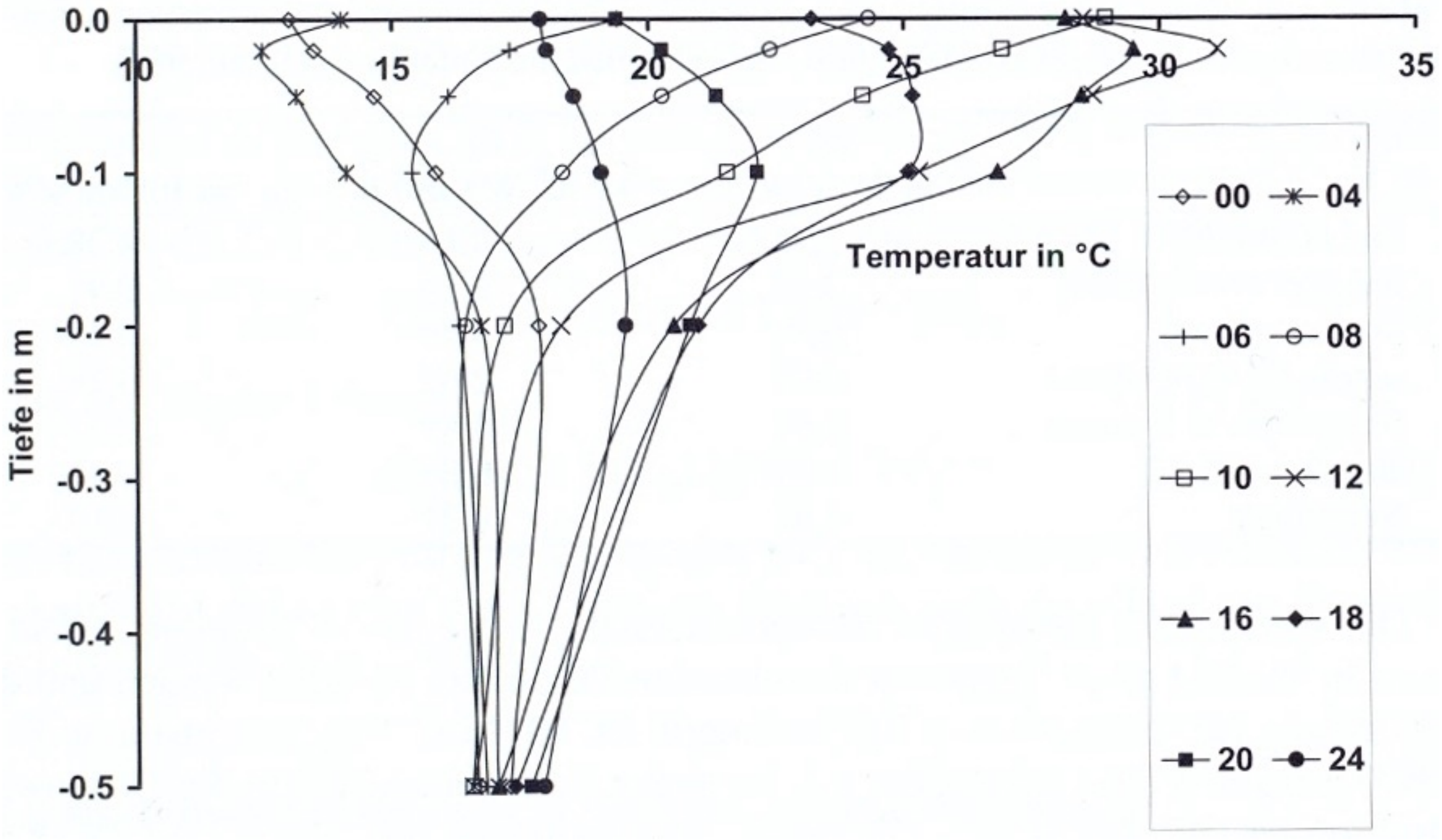
Ground heat flux

Ground flux represents the heat flux into the ground measured at the top of the soil.

Properties

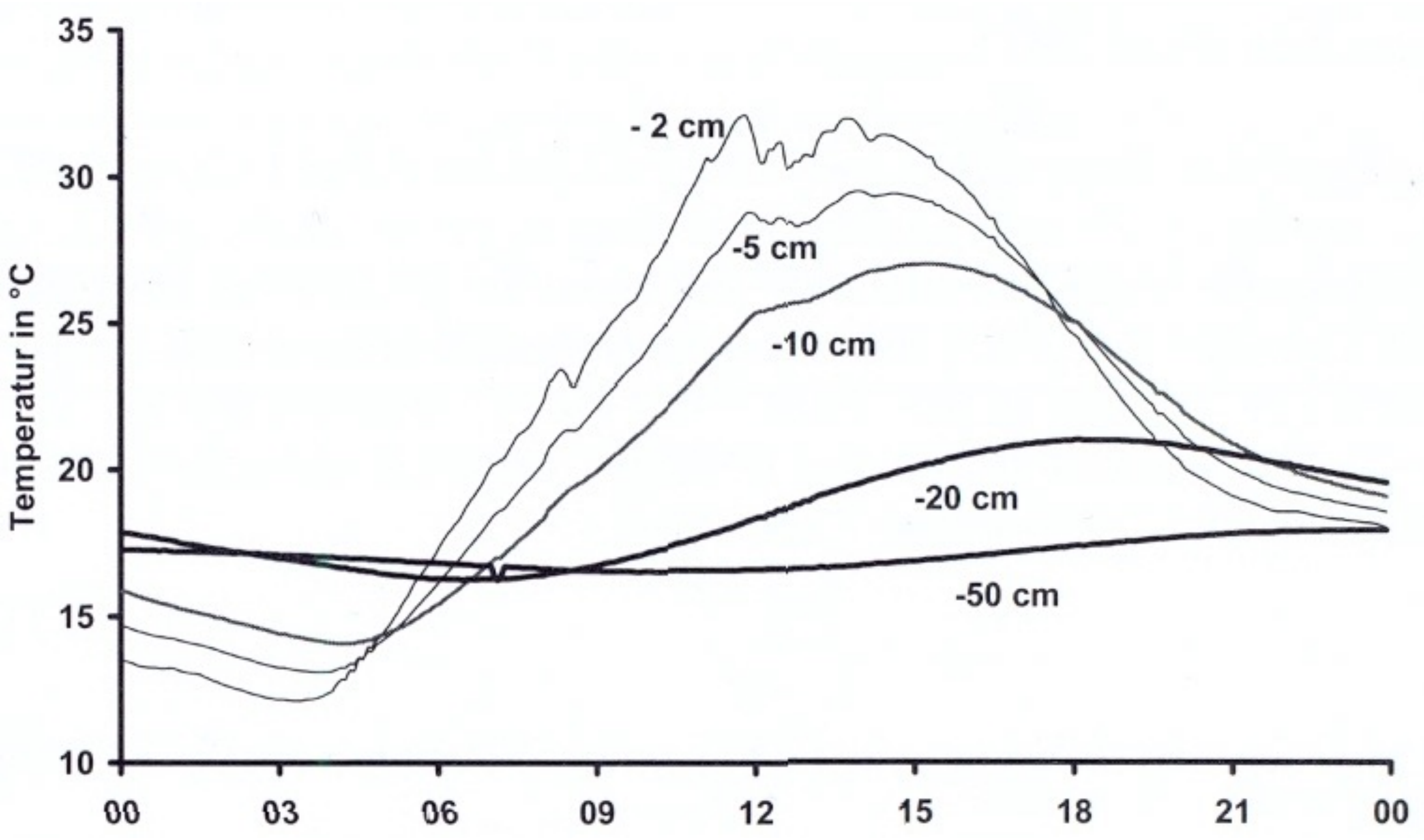
- Small but significant component of the surface energy budget
- The flux is related to the skin temperature
- If not measured, we need to parametrize it

Ground heat flux



Temperature profiles at indicated hours (Foken, 2008).

Ground heat flux



Daily cycle of soil temperatures in various depths (Foken, 2008).

Ground heat flux

Ground heat flux: $$Q_G = -k_g \frac{\partial T}{\partial z}$$ (Fourier's law)

Temporal temperature change: $$\frac{\partial T}{\partial t} = -\left(\frac{1}{c_g}\right)\frac{\partial Q_g}{\partial z}$$

Heat conductivity equation: $$\frac{\partial T}{\partial t} = \nu_g \frac{\partial^2 T}{\partial z^2}$$

with k_g *: Thermal molecular conductivity [W m^{-1} K^{-1}]*
c_g *: Soil heat capacity [Wm^{-3} K^{-1}]*
ν_g *: Soil thermal diffusivity ($\nu_g = k_g/c_g$) [m^2 s^{-1}]*

Molecular Conductivity, Heat Capacity and Thermal Diffusivity

Surface	Thermal molecular conductivity k_g in [$Wm^{-1}\ K^{-1}$]	Soil heat capacity C_g in 10^6 [$Wm^{-3}\ K^{-1}$]	Soil thermal diffusivity ν_g in 10^{-6} [$m^2 s^{-1}$]
Granit	2.73	2.13	1.28
Wet sand	2.51	2.76	0.91
Dry sand	0.30	1.24	0.24
Sandy clay	0.92	2.42	0.38
Swamp	0.89	3.89	0.23
Old snow	0.34	0.84	0.40
Fresh snow	0.02	0.21	0.10
Pure ice	2.10	2.09	1.09

Ground heat flux

Amplitude Change with Depth

$$\Delta T(z) = \Delta T_{surface} \cdot \exp\left[-z\left(\frac{\pi}{\nu_g P}\right)^{0.5}\right]$$

Phase Shift

$$\Delta t = \left(\frac{\Delta z}{2}\right)\left[\frac{P}{\pi \cdot \nu_G}\right]^{0.5}$$

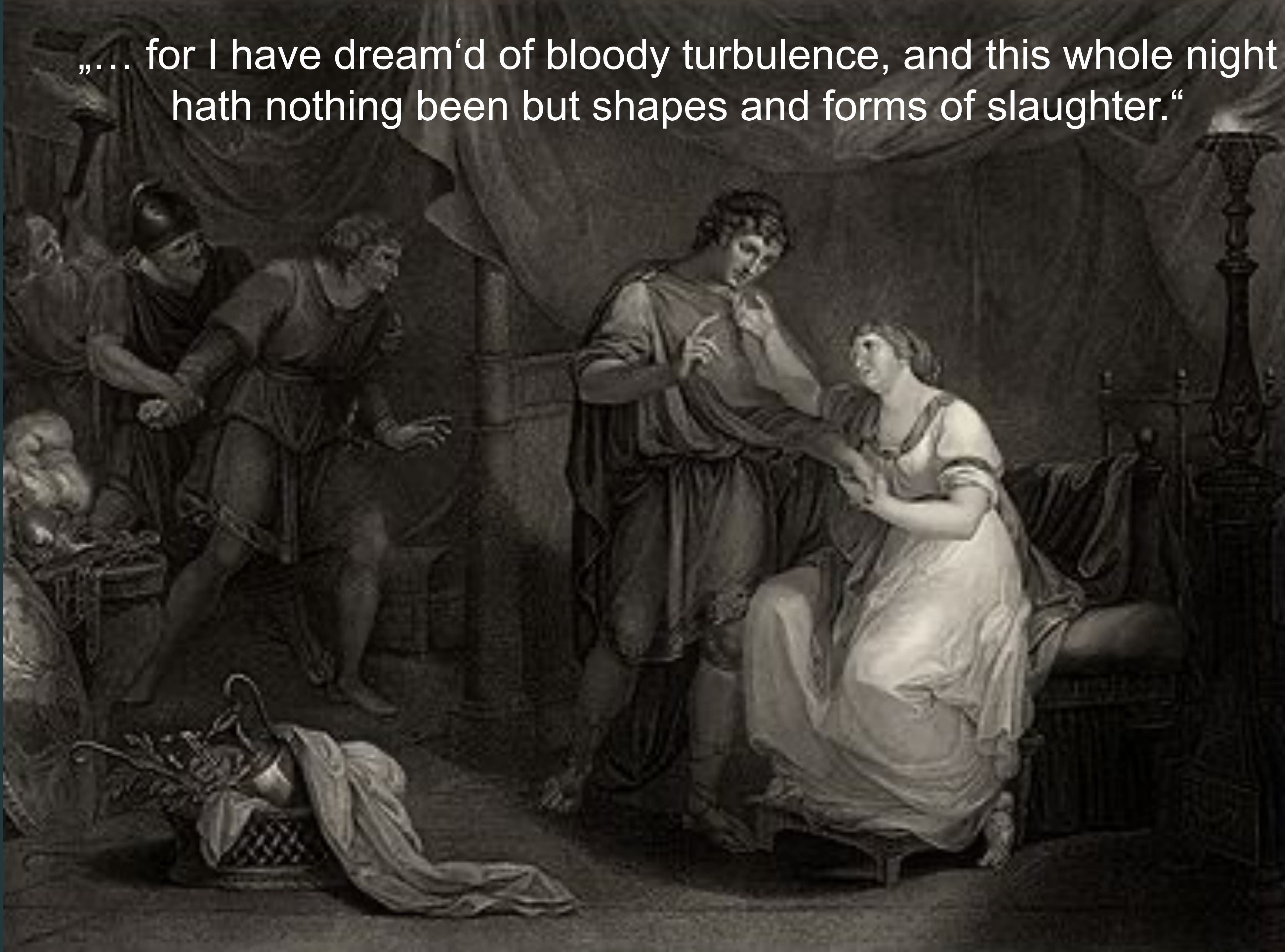
with ΔT : Amplitude of wave
P : Period of the cycle (daily or annual)

Parametrization turbulent surface fluxes

$$Q_S^* = Q_H + Q_E + Q_G + \Delta Q_s$$

William Shakespeare in „Troilus and Cressida“

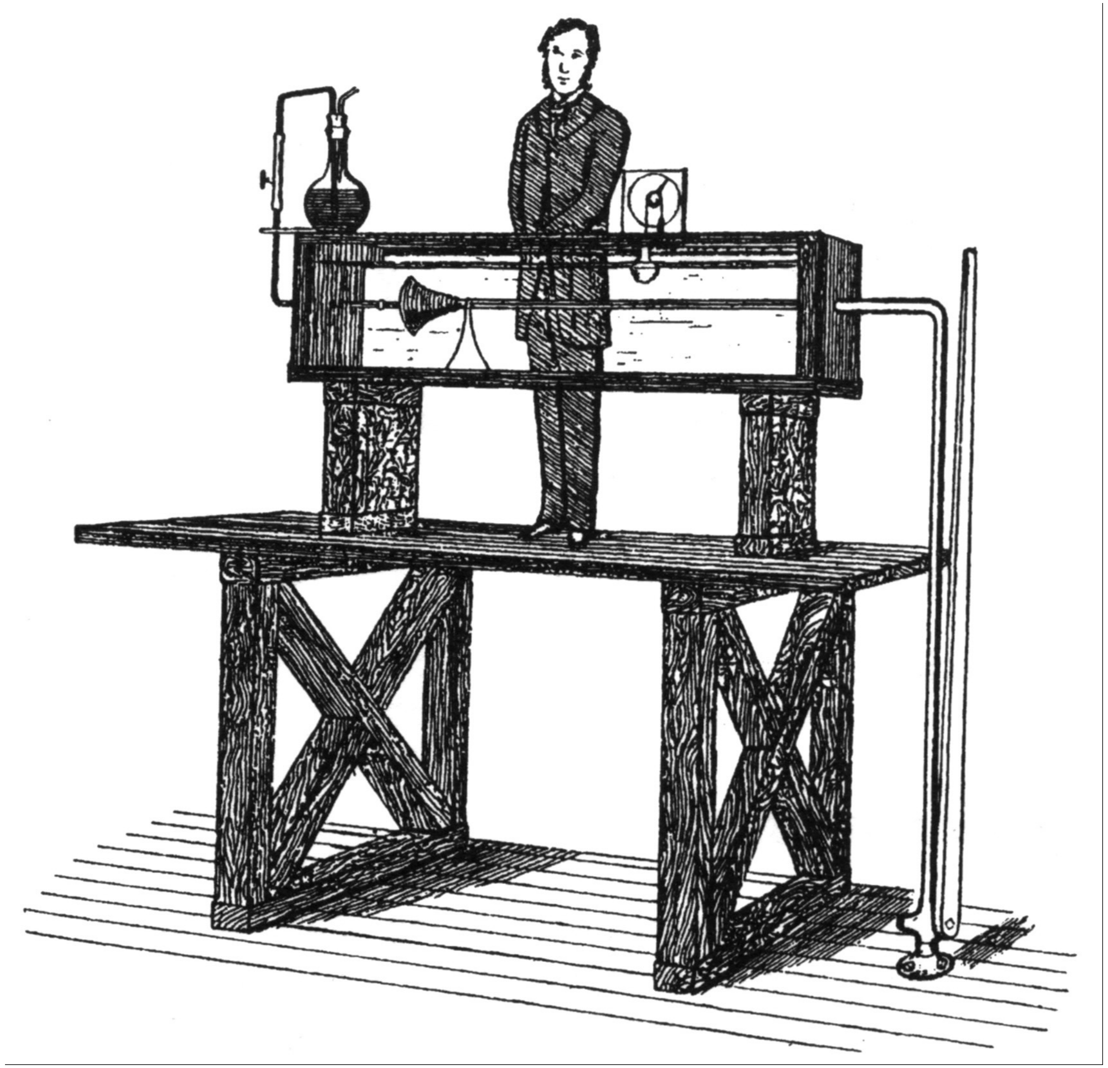
„… for I have dream‘d of bloody turbulence, and this whole night hath nothing been but shapes and forms of slaughter.“



Whirlpool of water from Leonardo Da Vinci (Anno 1513)

The great wave of Kanagawa from Katsushika Hokusai (ca. 1830).

Osborne Reynolds (1842-1912)

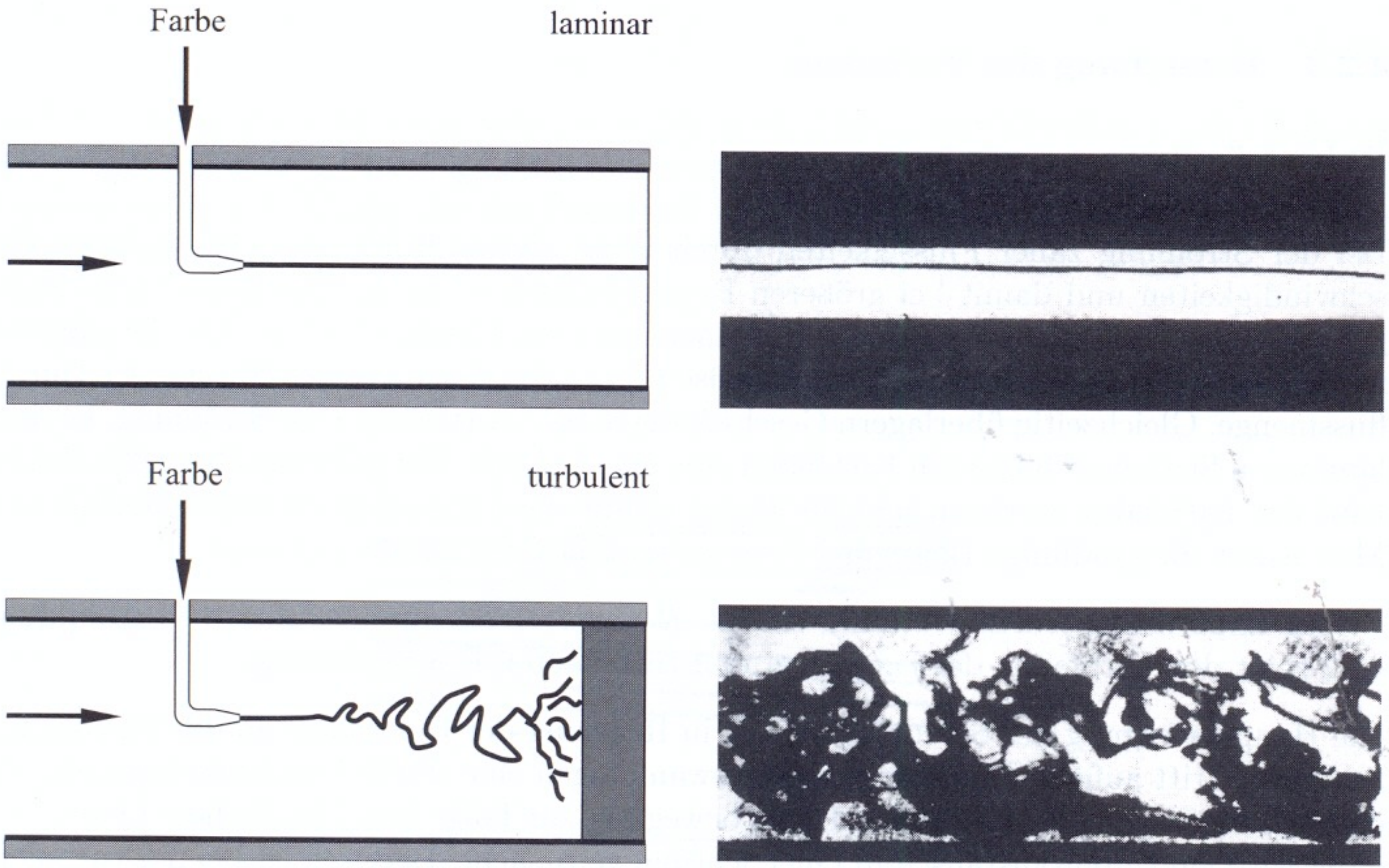


Reynolds experiment from 1883

“The equations of motion had been subjected to such close scrunity, particularly by Professor Stokes, that there was small chance to discover anything new or faulty in them. It seemed to me possible, however, that they might contain evidence which had been overlooked.”

“… either the elements of the fluid follow on another along lines of motion which lead in the most direct manner to their destinations, or they eddy about in sinous paths the most indirect possible."

Reynolds Experiment



Laminar and turbulent channel flow

Source: Oertel, 2008

Reynolds Experiment

Laminar and turbulent channel flow

Characteristics of turbulent flow

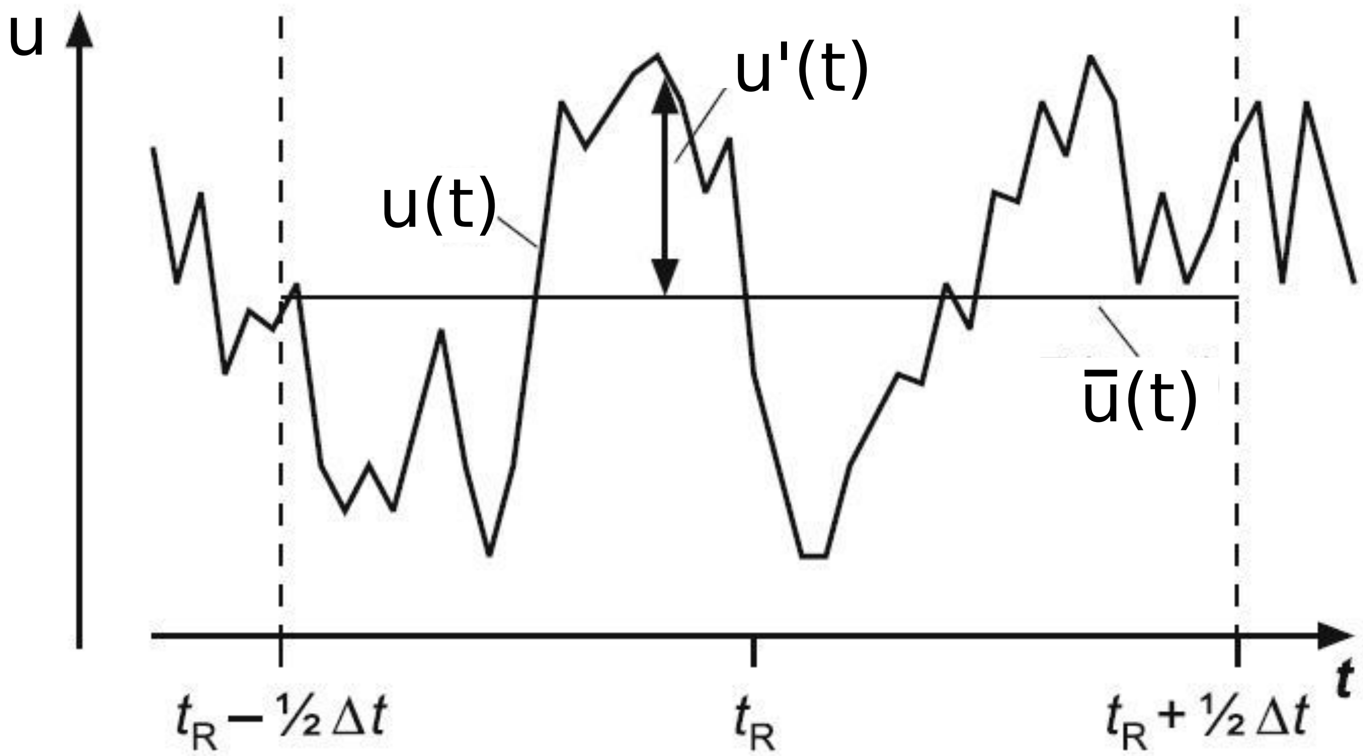
- The flows are rotational and three dimensional (vorticity fluctuation are therefore important)
- Flows are dissipative, so that energy must be supplied to maintain the turbulence
- The fluid motions are unpredictable in detail
- The rates of transfer and mixing are several orders of magnitude greater than the rate of molecular diffusion

Important Concepts

- Turbulence is stationary (statistical invariance with respect to time)
- Turbulence is homogeneous (statistically invariant to translation in space)
- Turbulence is isotropic (statistically independent of translation, rotation and reflection of the spatial axes)
- Turbulence is ergodic, if homogeneous and stationary.

Reynolds decomposition

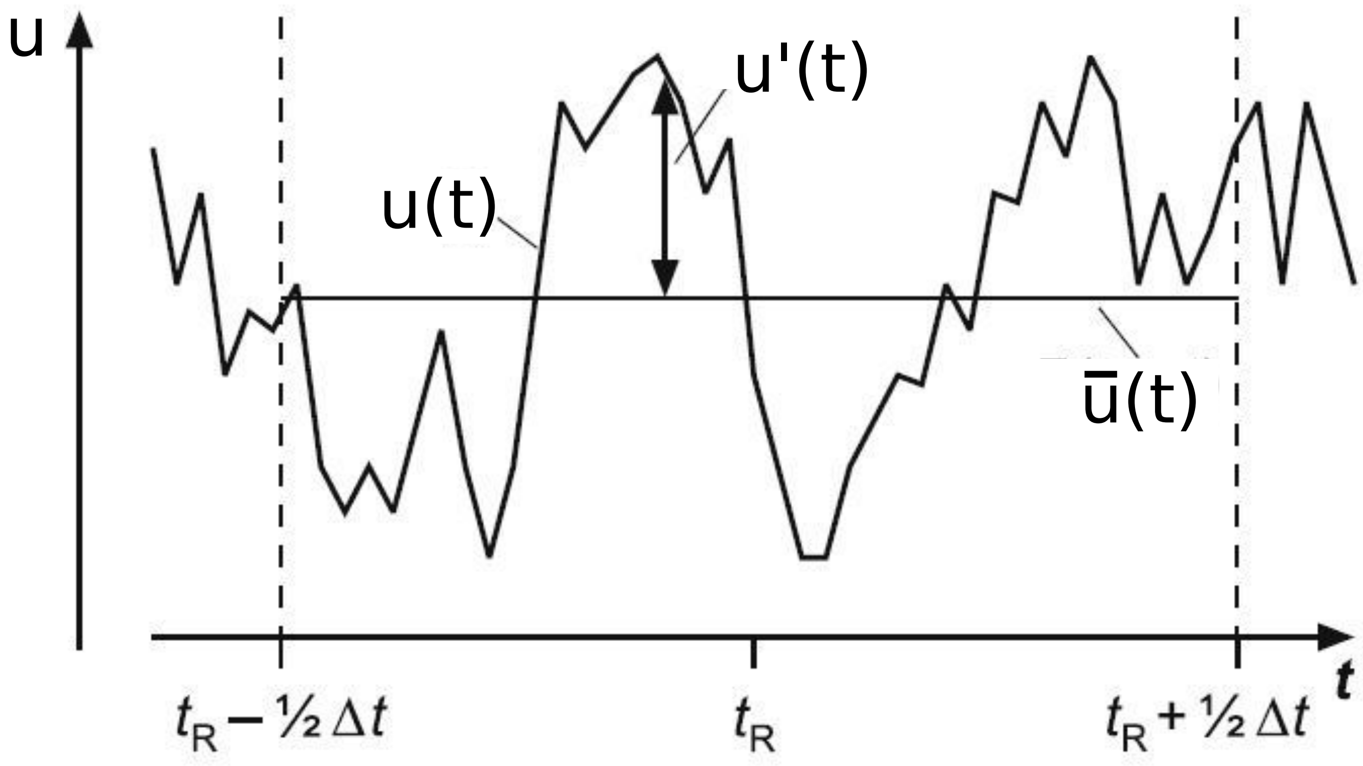
$$\boxed{u(t) = \bar{u}(t) + u'(t)}$$



Decomposition of a time series, showing the gusts $a'(t)$ of the actual instantaneous wind, $a(t)$, from the local mean, $\bar{a}(t_R, \delta t)$.

Source: modifed after Krauss, 2008

Reynolds decomposition



$$\boxed{\bar{u} = \overline{(\bar{u} + u')} = \bar{\bar{u}} + \overline{u'} = \bar{u} + \overline{u'}}$$

This is only true, if $\overline{u'} = 0$

Source: modifed after Krauss, 2008

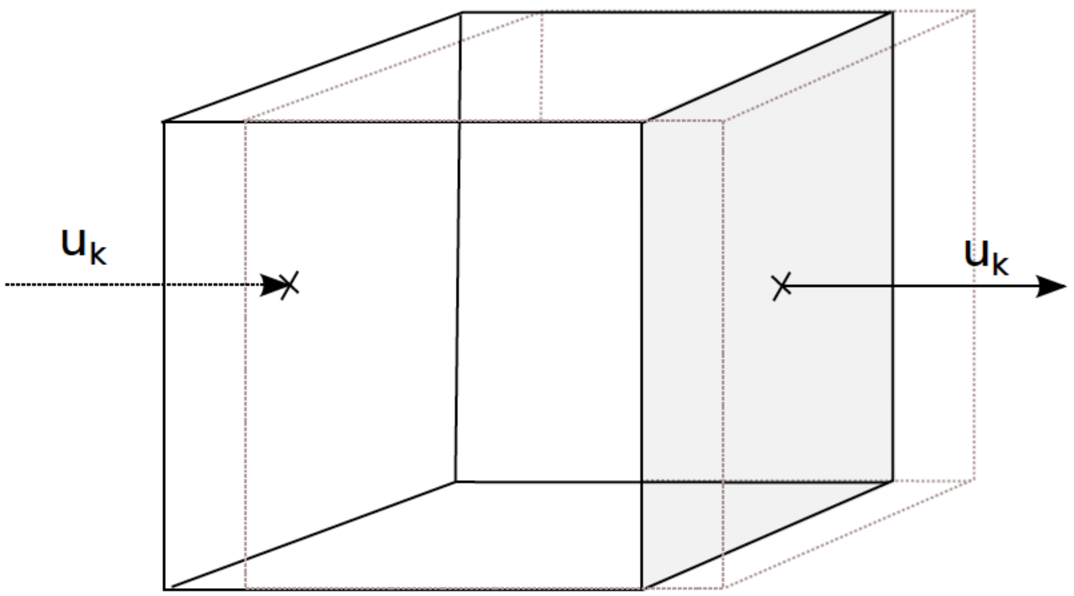
Properties of air masses

Air parcel contains properties, e.g. momentum, sensible heat, humidity, tracer concentration.

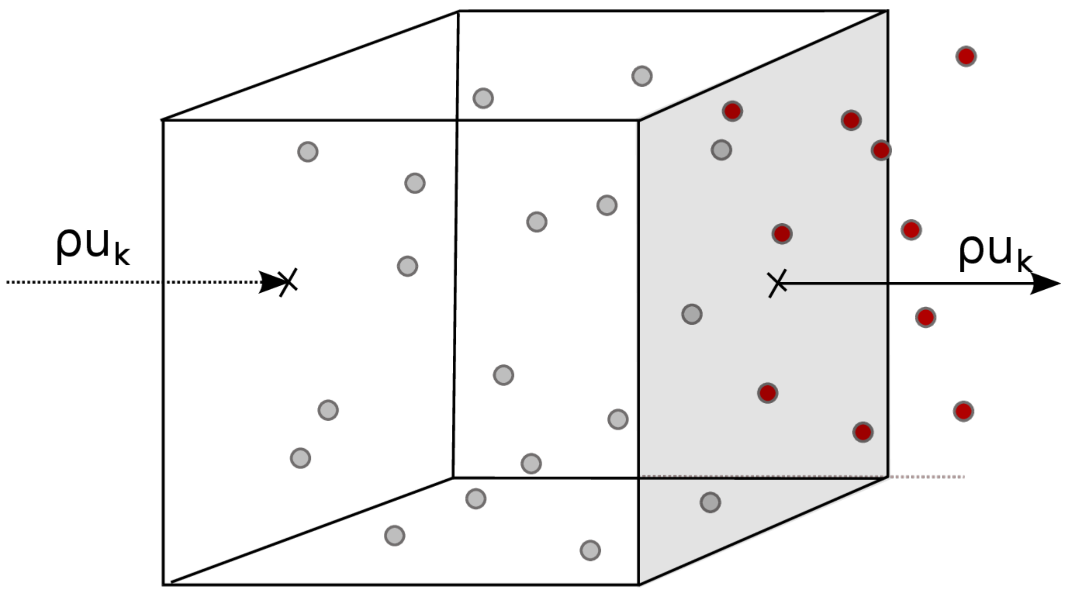
- ▷ **Momentum** [kg m s^{-1}] is defined as the product of mass m and velocity $\vec{v}$.
- ▷ **Sensible heat** [J kg^{-1}] is given as $c_p\theta$.
- ▷ **Specific humidity** [kg kg^{-1}], ratio between mass of vapor and mass of wet air.

➡ *Transport of specific properties are always related to the transport of air mass (mass flux).*

Fluxes



Schematic sketch of volume flux



Schematic sketch of mass flux

Fluxes	**Units**	**Description**
u_k	$ms^{-1} = m^3 m^{-2} s^{-1}$	Volume flux
ρu_k	$kgm^{-2}s^{-1}$	Mass flux
$\rho u_k e$	$kgm^{-2}s^{-1} \cdot \varepsilon \cdot kg^{-1} = \varepsilon m^{-2}s^{-1}$	Property flux

Table 1: Comparison of different fluxes, with k as the component of the wind speed ($u_1 = u$, $u_2 = v$, $u_3 = w$) and ε any property. The dry air density is given by ρ.

Reynolds decomposition of fluxes

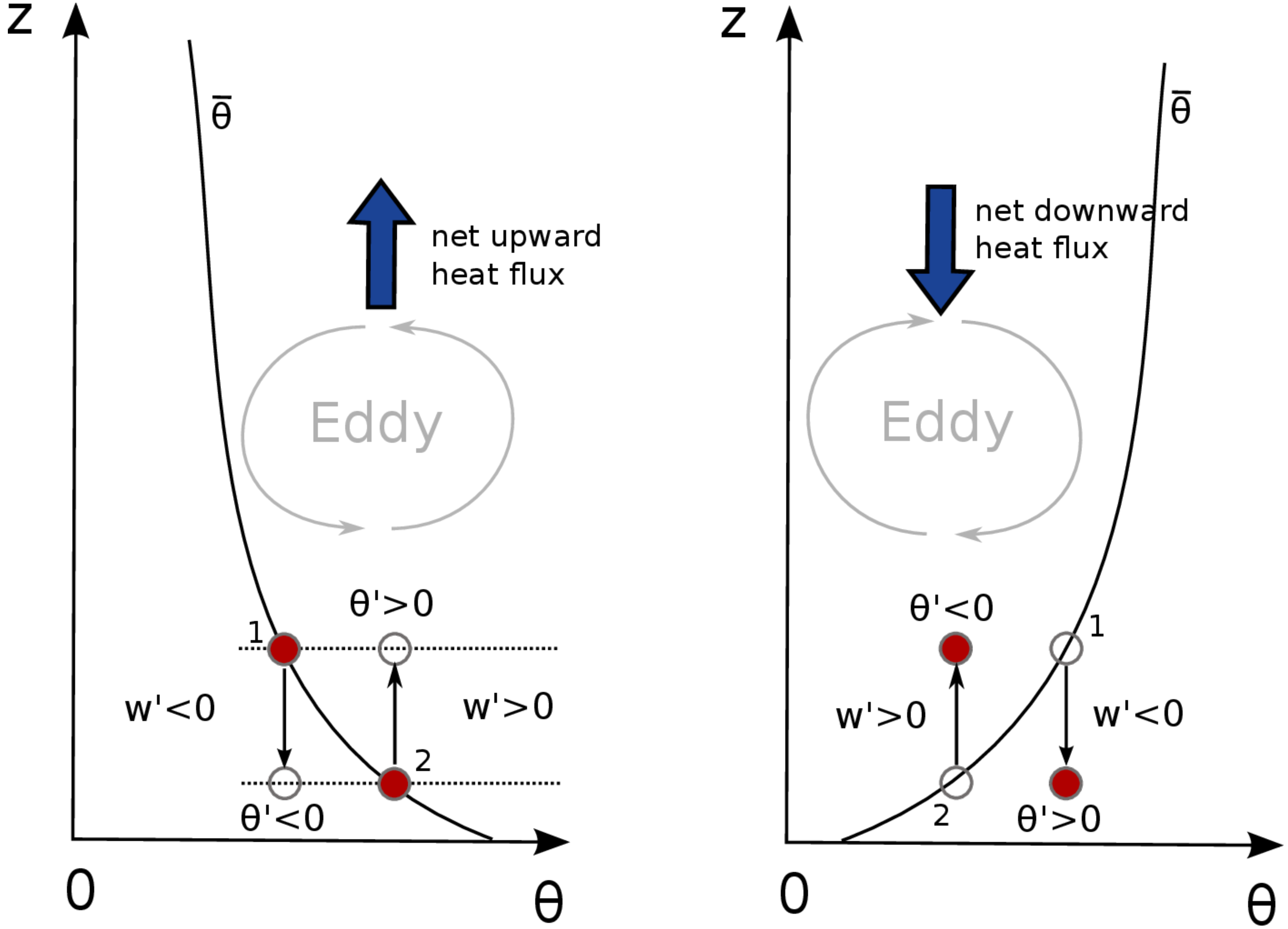
$$\overline{F_k^e} = \rho\overline{u_k e} = \rho \cdot \overline{(\overline{u_k} + u_k')(\bar{e} + e')}$$

$$= \rho\overline{u_k}\overline{e} + \rho\overline{u_k' e'}$$

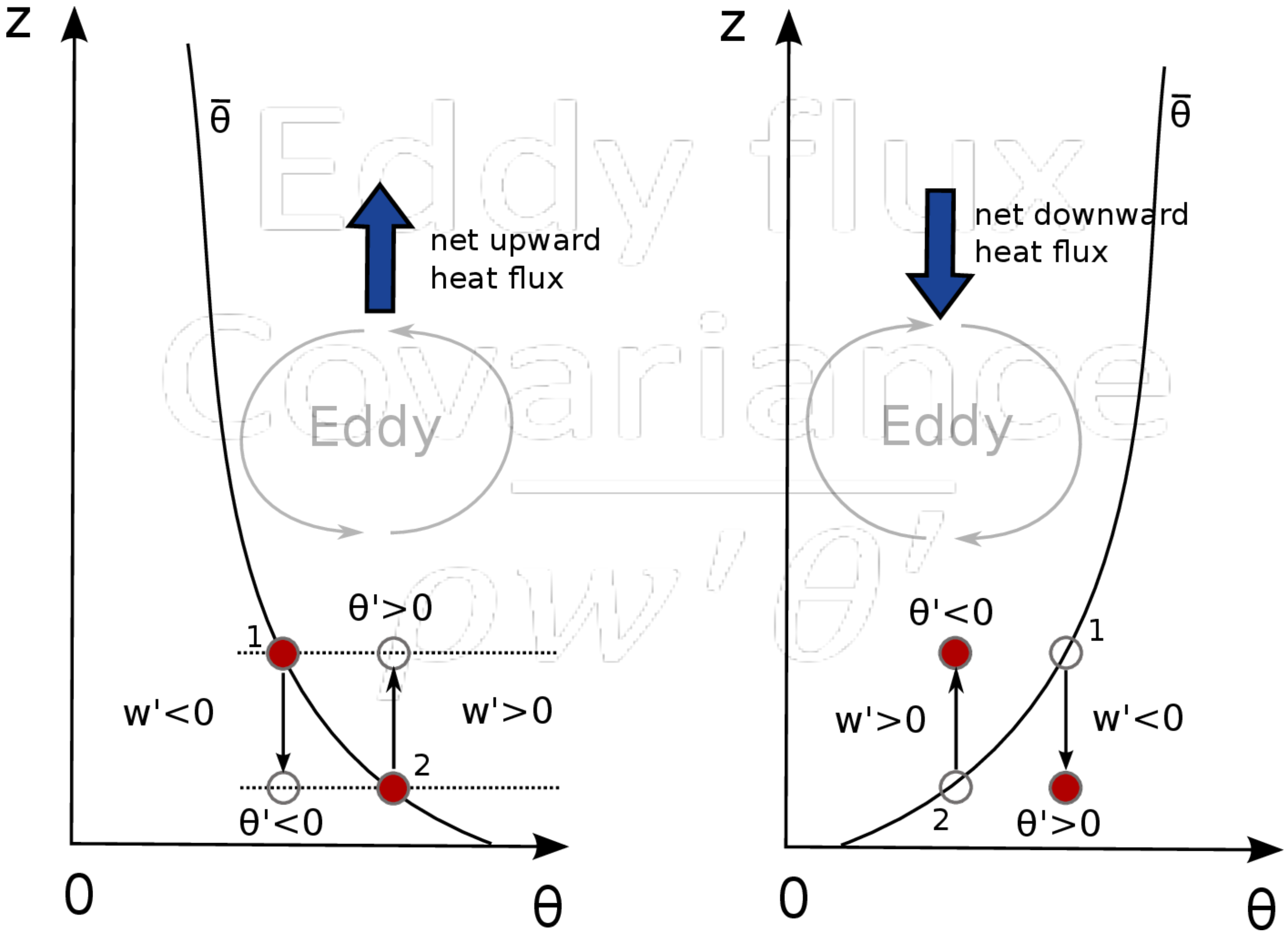
while $\rho\overline{u_k}\overline{e}$ is the **mean transport**

$\rho\overline{u_k' e'}$ is the **turbulent transport**

Eddy flux
Covariance
$$\overline{\rho w'\theta'}$$



Eddy flux Covariance $\overline{\rho w'\theta'}$



w'	$+$	$+$	-	-
θ'	$+$	-	$+$	-
	warm air from below	cold air from below	warm air from above	cold air from above
$\overline{w'\theta'}$	$+$	-	-	$+$

Overview: Kinematic fluxes

Vertical kinematic eddy sensible heat flux	$= \overline{w'\theta'}$	[K m s^{-1}]
Vertical kinematic eddy moisture flux	$= \overline{w'q'}$	[kg kg^{-1} m s^{-1}]
Vertical kinematic eddy flux of u-momentum	$= \overline{u'w'}$	[m^2 s^{-2}]

Turbulent heat fluxes

Vertical turbulent sensible heat flux

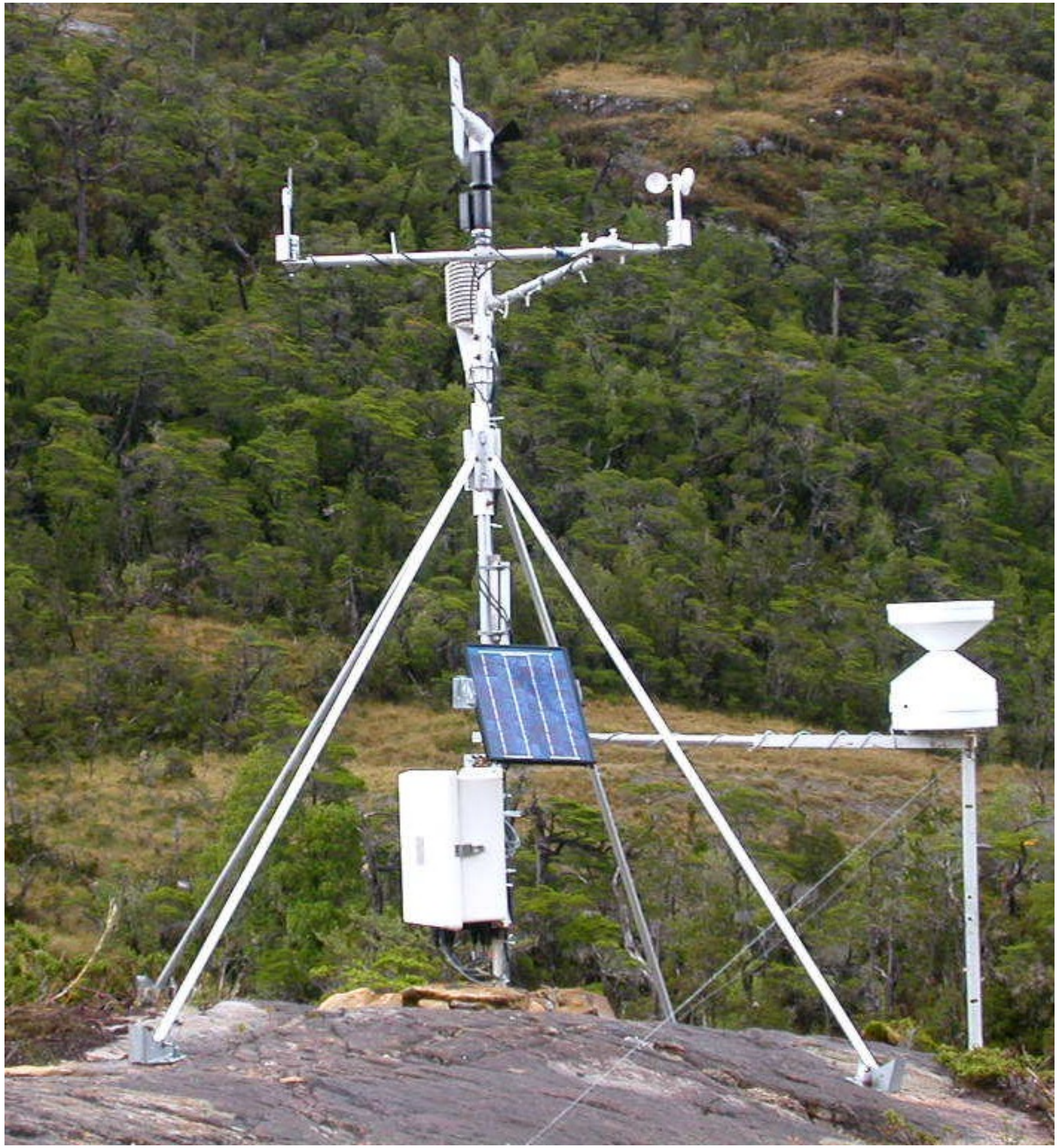
$$Q_H = c_p \cdot \rho \cdot \overline{w'\theta'} \qquad [\mathrm{W\ m^{-2}}]$$

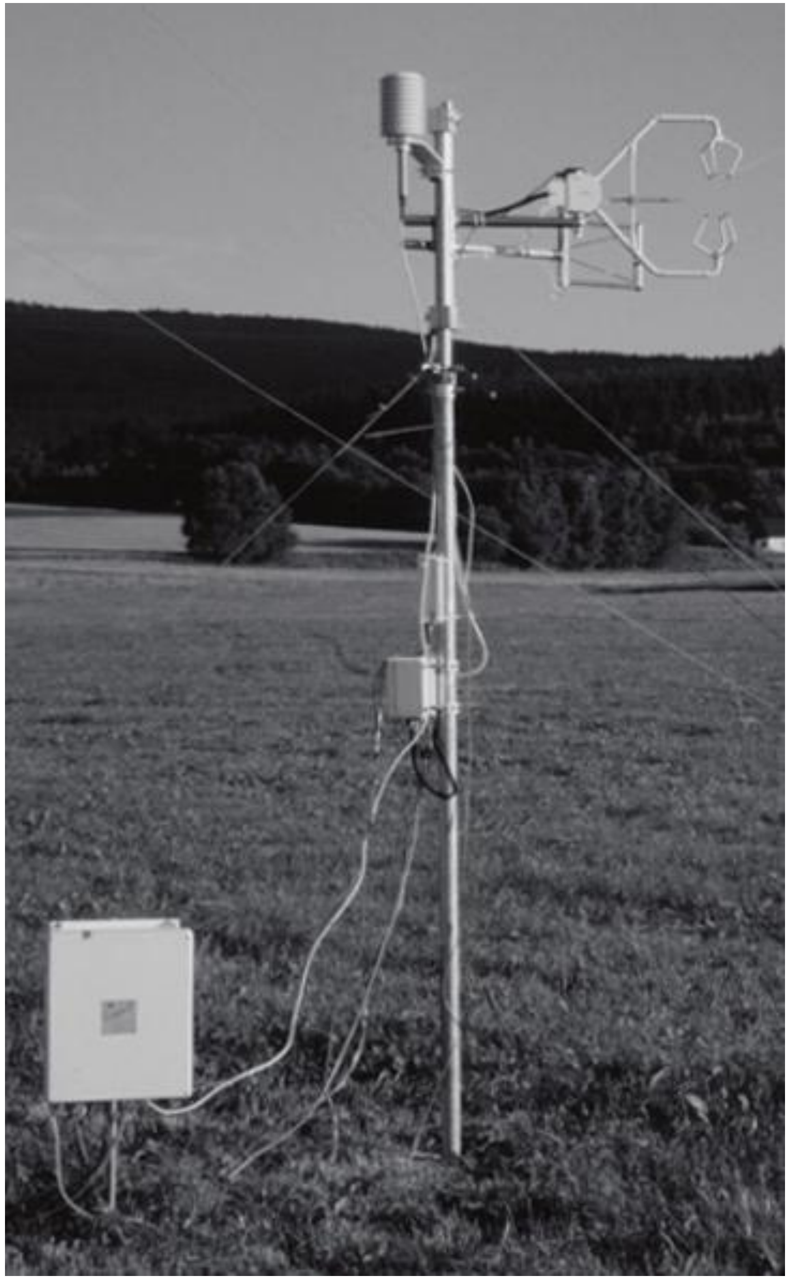
Vertical turbulent latent heat flux

$$Q_E = L \cdot \rho \cdot \overline{w'q'} \qquad [\mathrm{W\ m^{-2}}]$$

with L : latent heat of vaporization of water 334 [kJ kg^{-1}]
c_p: specific heat of constant pressure for moist air 1.004 [kJ kg^{-1} K^{-1}]

Problem: Covariances





K-theory First order
(1. Method)

Parametrization of turbulent fluxes as

$$\overline{u'_k e'} = -K \frac{\partial \bar{e}}{\partial x_k}$$

with K the eddy diffusivity [$m^2\ s^{-1}$]

Different K-values are associated with different variables, K_m for momentum, K_H and K_E for heat and latent heat fluxes.

$$K_H = K_E = 1.35 \cdot K_m$$ for neutral conditions

Turbulent fluxes

Vertical turbulent heat flux

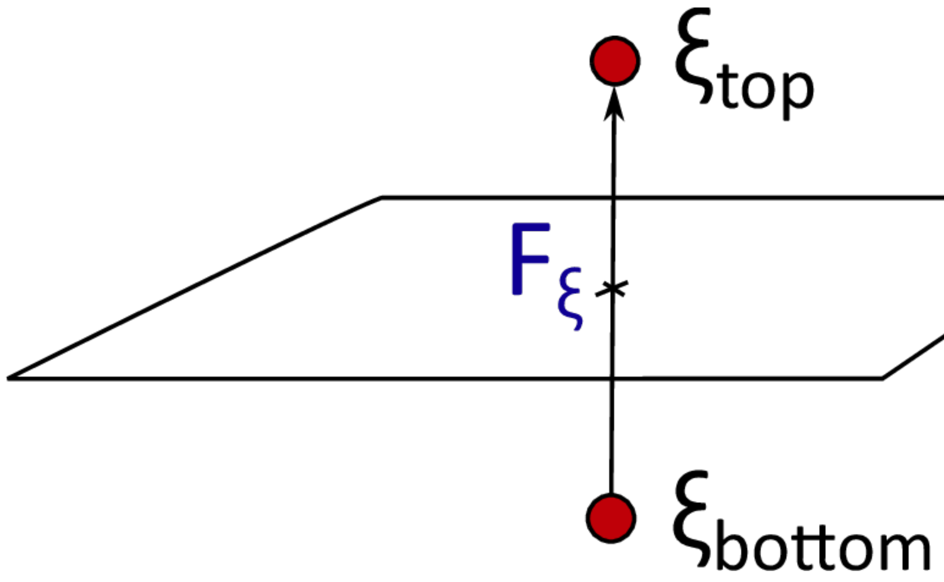
$$Q_H = c_p \cdot \rho \cdot \overline{w'\theta'} = c_p \cdot \rho \cdot K_H \frac{\partial \overline{\theta}}{\partial z} \quad [W\ m^{-2}]$$

Vertical turbulent latent heat flux

$$Q_E = L \cdot \rho \cdot \overline{w'q'} = L \cdot \rho \cdot K_E \frac{\partial \overline{q}}{\partial z} \quad [W\ m^{-2}]$$

with L : latent heat of vaporization of water 334 [KJ kg^{-1}]
c_p : specific heat at constant pressure for moist air 1006 [KJ kg^{-1} K^{-1}]

Parametrization turbulent surface fluxes (2. Method)



$$F_\zeta = -U_T \cdot (\zeta_{top} - \zeta_{bottom})$$

with U_T the transport velocity [m s^{-1}], and ζ any variable

The transport velocity is usually parametrized as a function of some measure of turbulence:

$$U_T = C_D \cdot \overline{U}$$

with C_D the bulk transfer coefficient

Parametrization turbulent surface fluxes

$$\boxed{\overline{w'\theta'} = -C_H \cdot U \cdot (T_z - T_0)}$$

$$\boxed{\overline{w'q'} = -C_E \cdot U \cdot (q_z - q_0)}$$

with C_H and C_E the bulk transfer coefficient, typical values range from 1e-3 to 1e-5 (C_H is the so-called Stanton number, C_E is the so-called Dalton number)

Subscript z represents values in the air near the surface (2 m or 10 m).
Subscript 0 means the value in the top 1 mm of the soil or sea surface.

Parametrization turbulent surface fluxes (Bulk Approach)

Turbulent sensible heat flux

$$\boxed{H = \rho_a \cdot c_p \cdot C_h \cdot u \cdot (T_z - T_0)}$$

$$C_h = \frac{\kappa^2}{ln\left(\frac{z}{z_0}\right)^2}$$

Turbulent latent heat flux

$$C_E = \frac{\kappa^2}{ln\left(\frac{z}{z_0}\right)^2}$$

$$\boxed{E = \rho_a \cdot L \cdot C_E \cdot u \cdot (q_z - q_0)}$$

with ρ : air density [kg m-3]

c_p : specific heat of constant pressure for moist air 1006 [kJ kg^{-1} K^{-1}]

L : latent heat of vaporization of water 334 [kJ kg^{-1}]

Drag coeff., measurement height and surface roughness

Statically neutral conditions in the surface layer

$$C_{DN} = \kappa^2 \left[\ln\left(\frac{z}{z_0}\right) \right]^{-2}$$

often $C_{DN} = C_{HN} = C_{EN}$, and κ (0.4) the von Karman constant

Statically unstable/stable flows

$$C_D = \kappa^2 \left[\ln\left(\frac{z}{z_0}\right) - \psi_M(\zeta) \right]^{-2}$$

$$C_H = \kappa^2 \left[\ln\left(\frac{z}{z_0}\right) - \psi_M(\zeta) \right]^{-1} \left[\ln\left(\frac{z}{z_0}\right) - \psi_H(\zeta) \right]^{-1}$$

with

ψ_M, ψ_H	:	Stability correction terms
ζ	:	Measure of stability ($\zeta = z/L$, while L is the Obukhov-Length)
		It is usually assumed that $C_H = C_E$
z_0	:	Roughness length $[m]$

Bulk coefficients

The coefficients depend on

- Surface roughness (frictional skin drag, form drag, wave drag)
- Measurement height (dramatic impact on the value of the drag coefficient)
- Stability

Aerodynamic roughness length z_0

Properties

- Height where wind speed becomes zero
- Aerodynamic, because the only true determination of this parameter follows from measurements
- Given observations of wind speed at two or more heights, it is easy to solve for z_0 and u
- It is not equal to the height of the individual roughness elements
- It does not change with wind speed, stability or stress

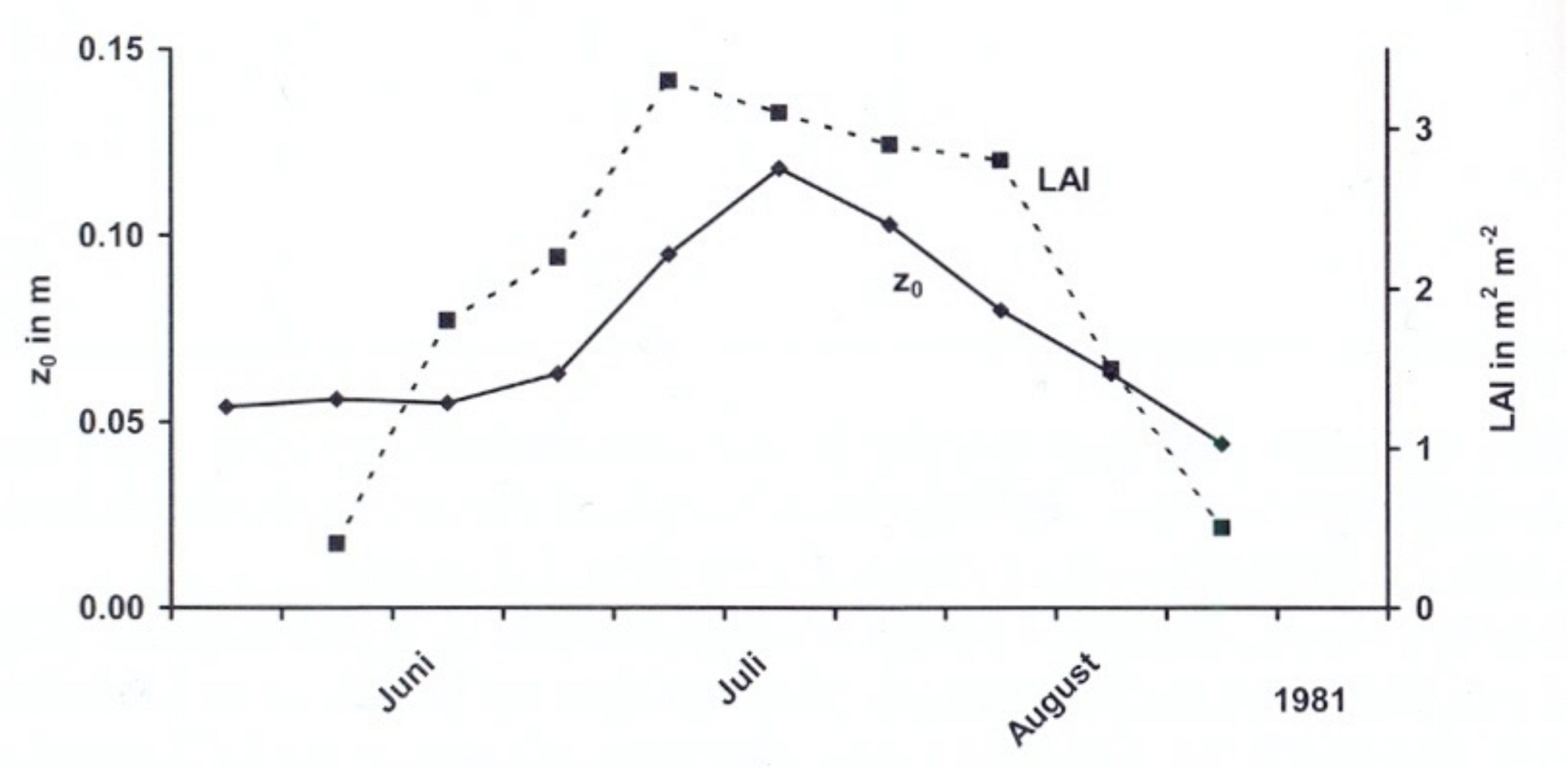
Aerodynamic roughness length z_0

$$Z_0 = \frac{1}{10} h \quad [\mathrm{m}]$$

with h the height of the roughness elements

Surface	z_0 **in m**
Snow without structure	$10^{-5} \ldots 10^{-4}$
Calm water surface	$10^{-5} \ldots 10^{-4}$
Sand desert	$10^{-4} \ldots 10^{-3}$
Short grass (2-5 cm)	$2 \cdot 10^{-3} \ldots 5 \cdot 10^{-3}$
Long grass (20-50 cm)	$2 \cdot 10^{-2} \ldots 5 \cdot 10^{-2}$
Corn field	$2 \cdot 10^{-2} \ldots 5 \cdot 10^{-2}$
Forest	$1 \ldots 5$

Aerodynamic roughness length z_0



Change of roughness height and the leaf area index (Foken, 2008).

Bowen-Ratio Method (3. Method)

Bowen-Ratio

$$Bo = \frac{Q_H}{Q_E} = \gamma \cdot \frac{\Delta\theta}{\Delta e}$$

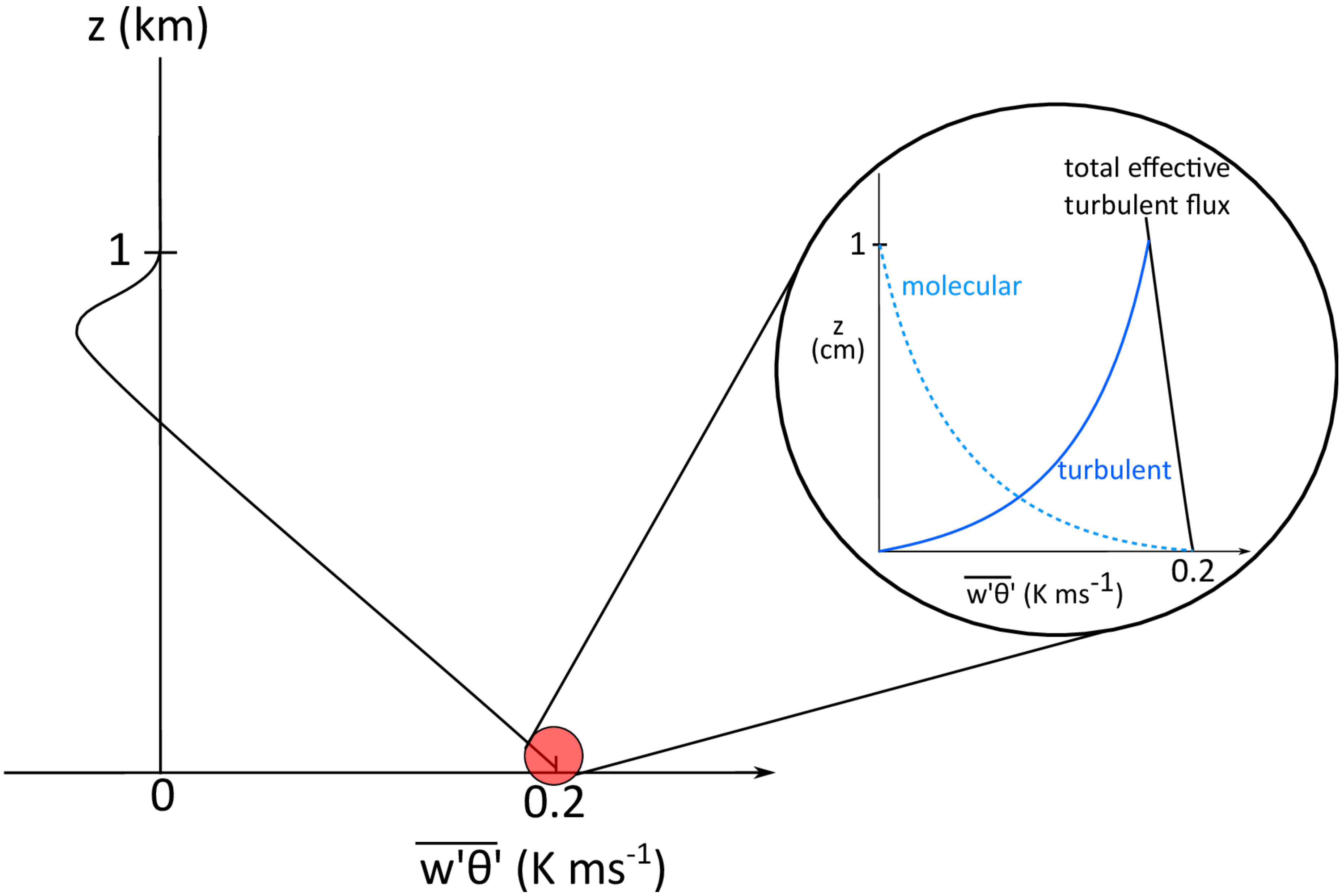
Heat fluxes

$$Q_H = (Q_s^* - Q_G)\frac{Bo}{1 + Bo}$$

$$Q_E = \frac{Q_S^* - Q_G}{1 + Bo}$$

with γ : Psychromatic constant 0.000667 $[Pa\ K^{-1}]$ for 1000 hPa and 20°C

Effective surface turbulent flux

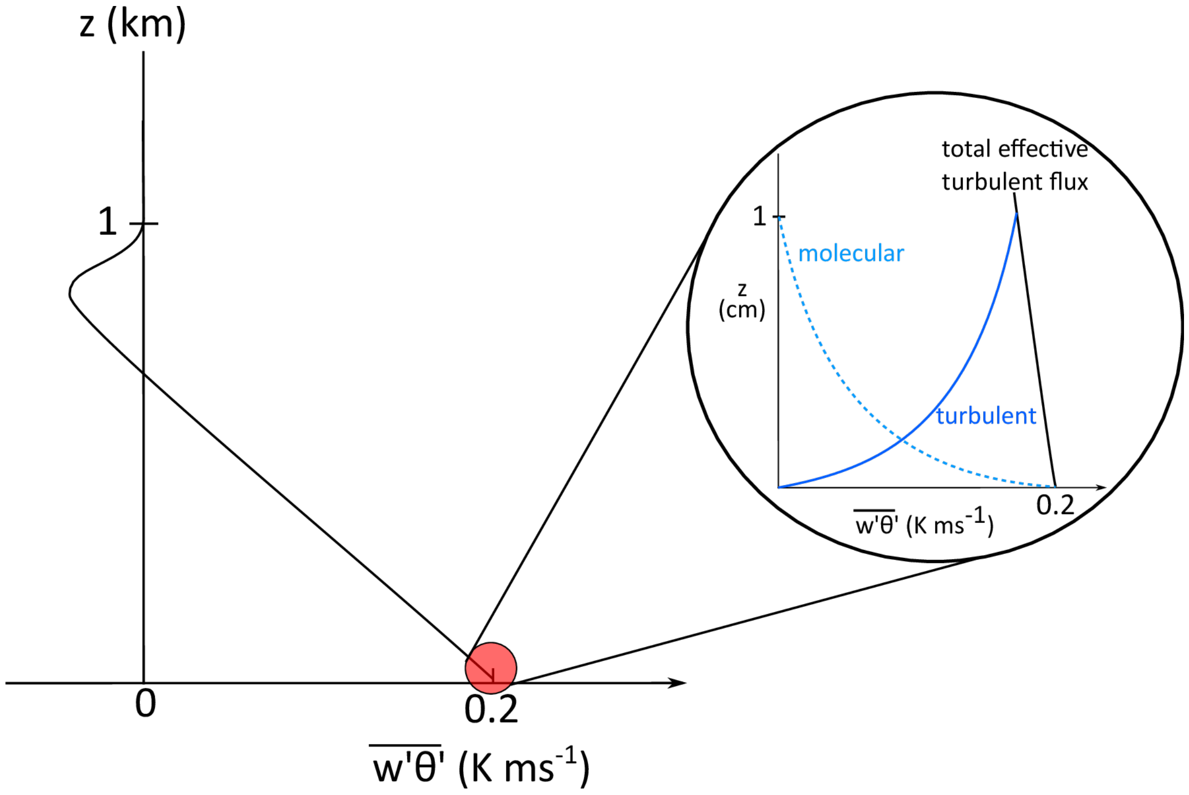


Effective surface turbulent flux

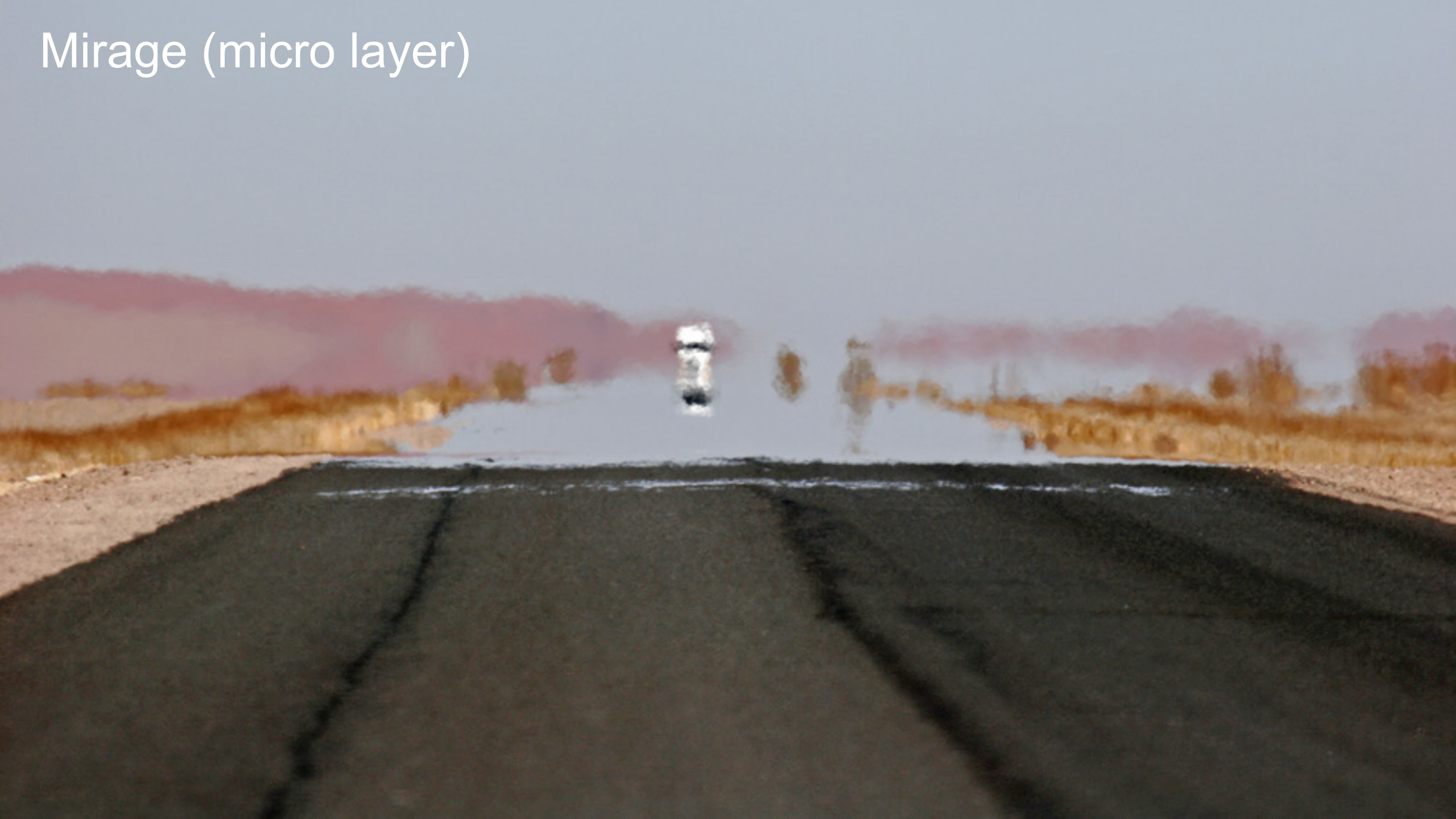
Molecular flux

$$Q_H = \nu_\theta \frac{\partial T}{\partial z} \qquad [\mathrm{W\ m^{-2}}]$$

with ν_θ the molecular thermal diffusivity ($2.5 \cdot 10^{-2}$ W m^{-1} K^{-1})

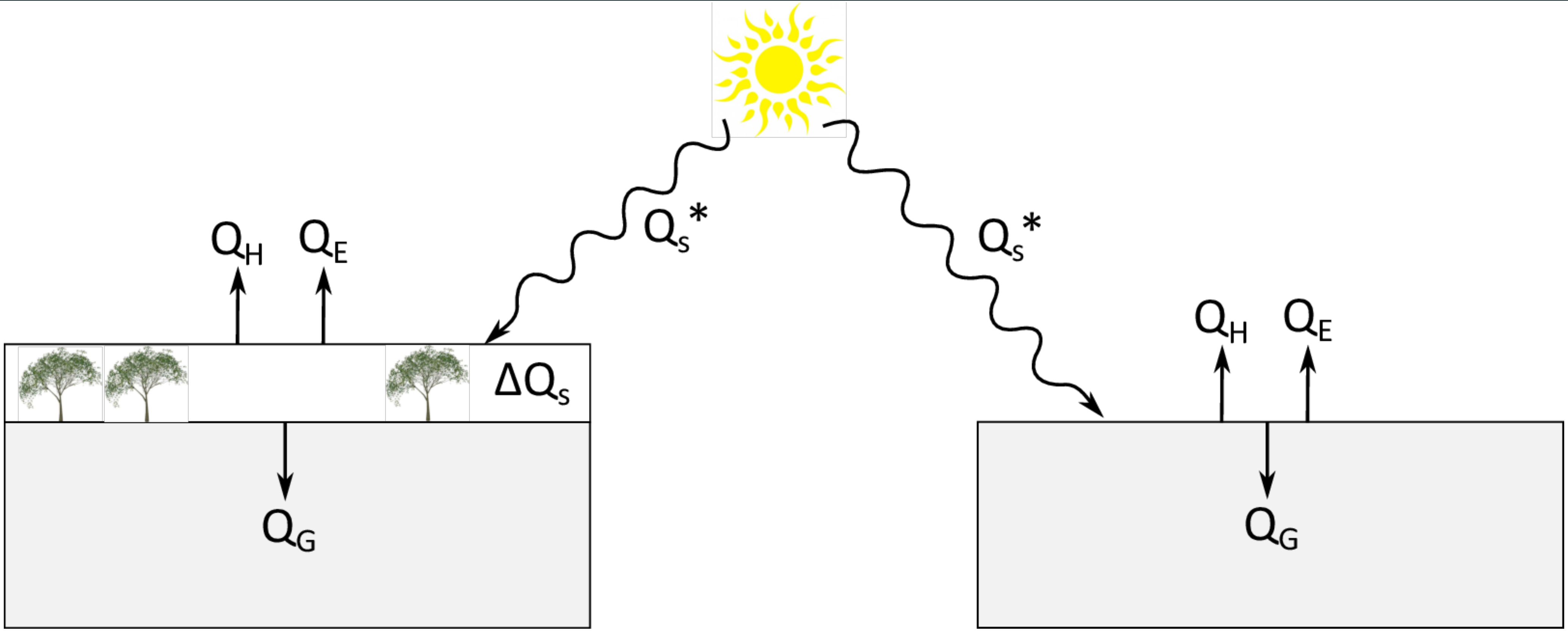


Mirage (micro layer)



Surface Energy Balance

Surface Energy Balance

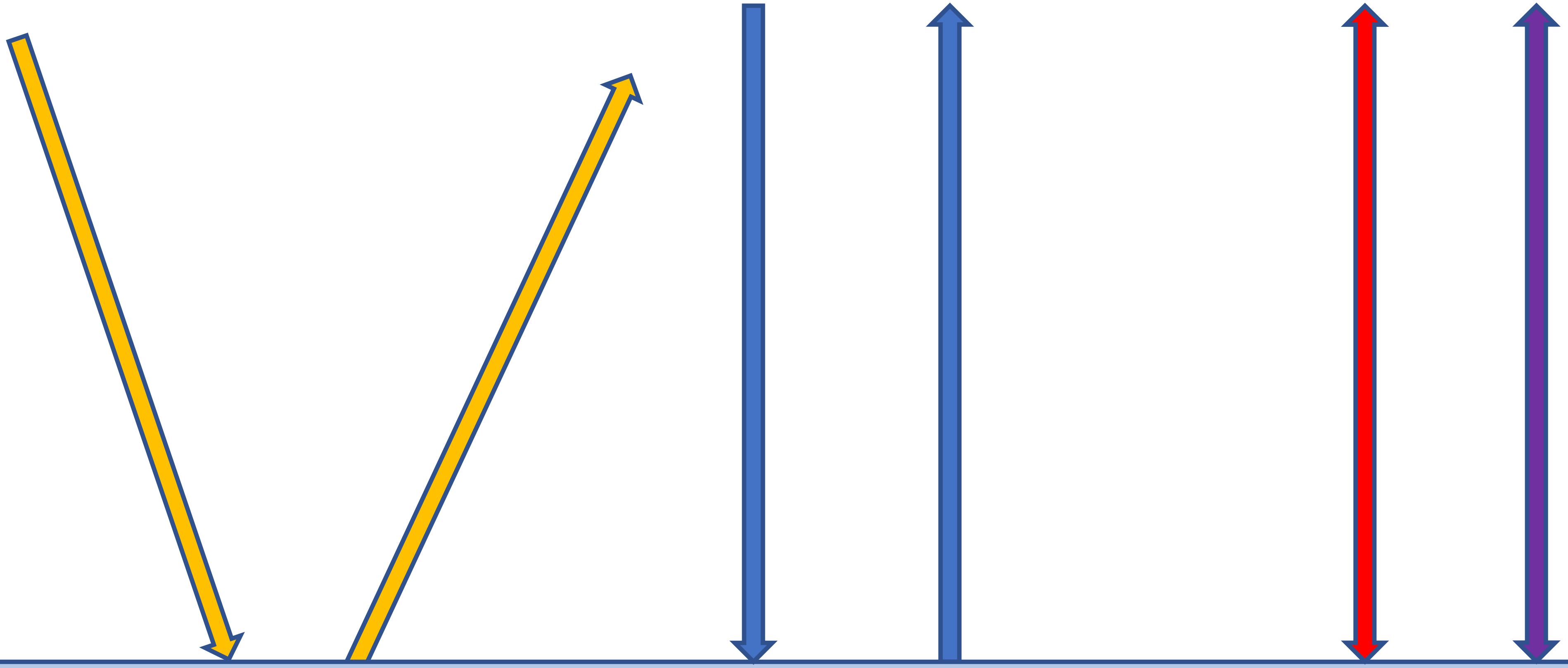


$$Q_S^* = Q_H + Q_E + Q_G + \Delta Q_s$$

with Q_S^* the net radiation at the surface, Q_H the sensible heat flux, Q_E the latent heat flux, Q_G the ground heat flux, and ΔQ_s the storage term (internal energy)

FLUXES AWAY FROM SURFACE ARE POSITIVE, EXCEPT FOR NET RADIATION

Surface Energy Balance model (SEB)



$$C_h = \frac{\kappa^2}{ln\left(\frac{z}{z_0}\right)^2}$$

$$C_E = \frac{\kappa^2}{ln\left(\frac{z}{z_0}\right)^2}$$

0 = SW + LW + H + E + G

$$SW = (1-\alpha)\cdot q_G$$

$$LW_d = \epsilon_s \cdot \sigma \cdot T_a^4$$

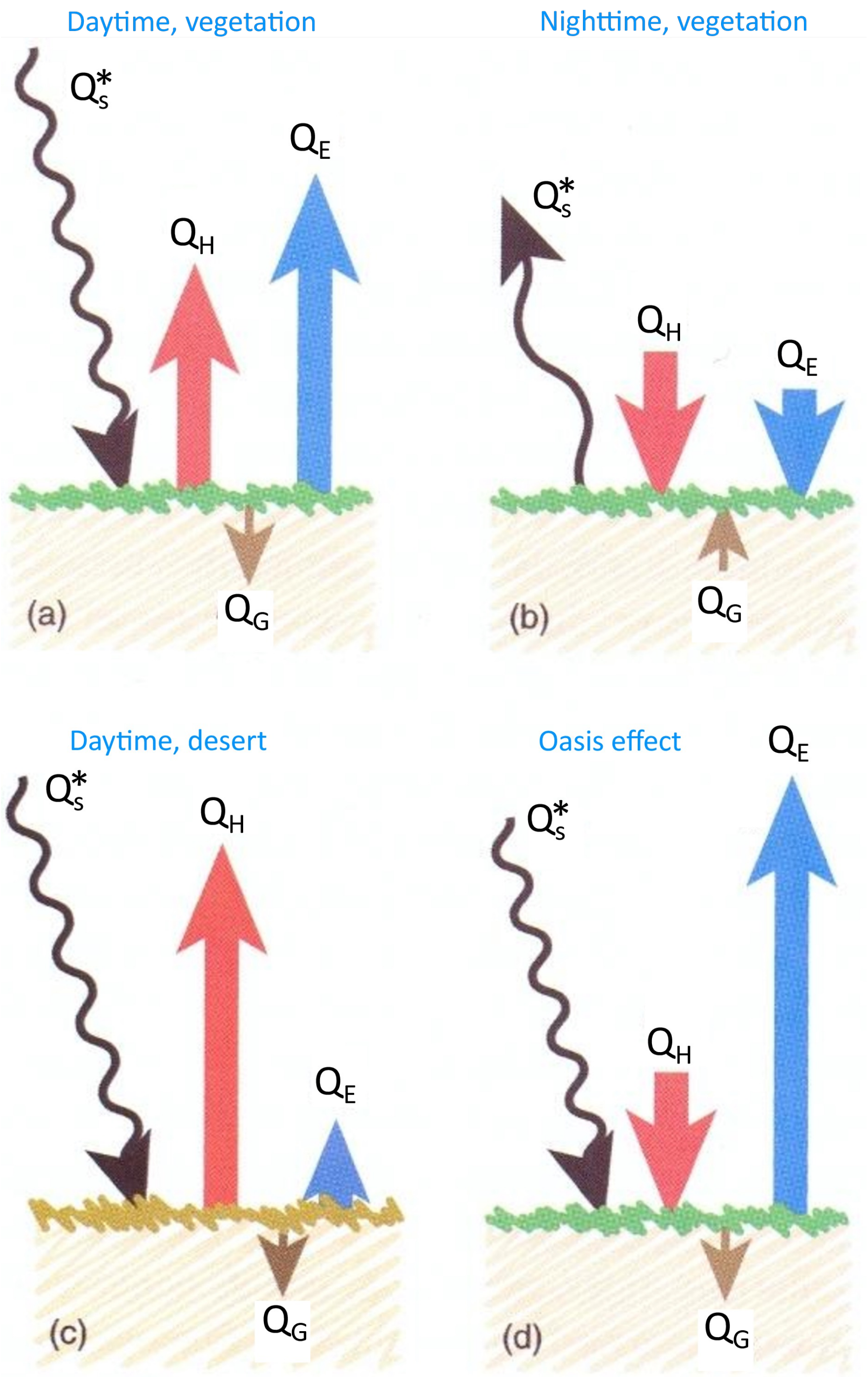
$$LW_u = \epsilon_a \cdot \sigma \cdot T_0^4$$

$$H = \rho_a \cdot c_p \cdot C_h \cdot u \cdot (T_z - T_0)$$

$$E = \rho_a \cdot L \cdot C_E \cdot u \cdot (q_z - q_0)$$

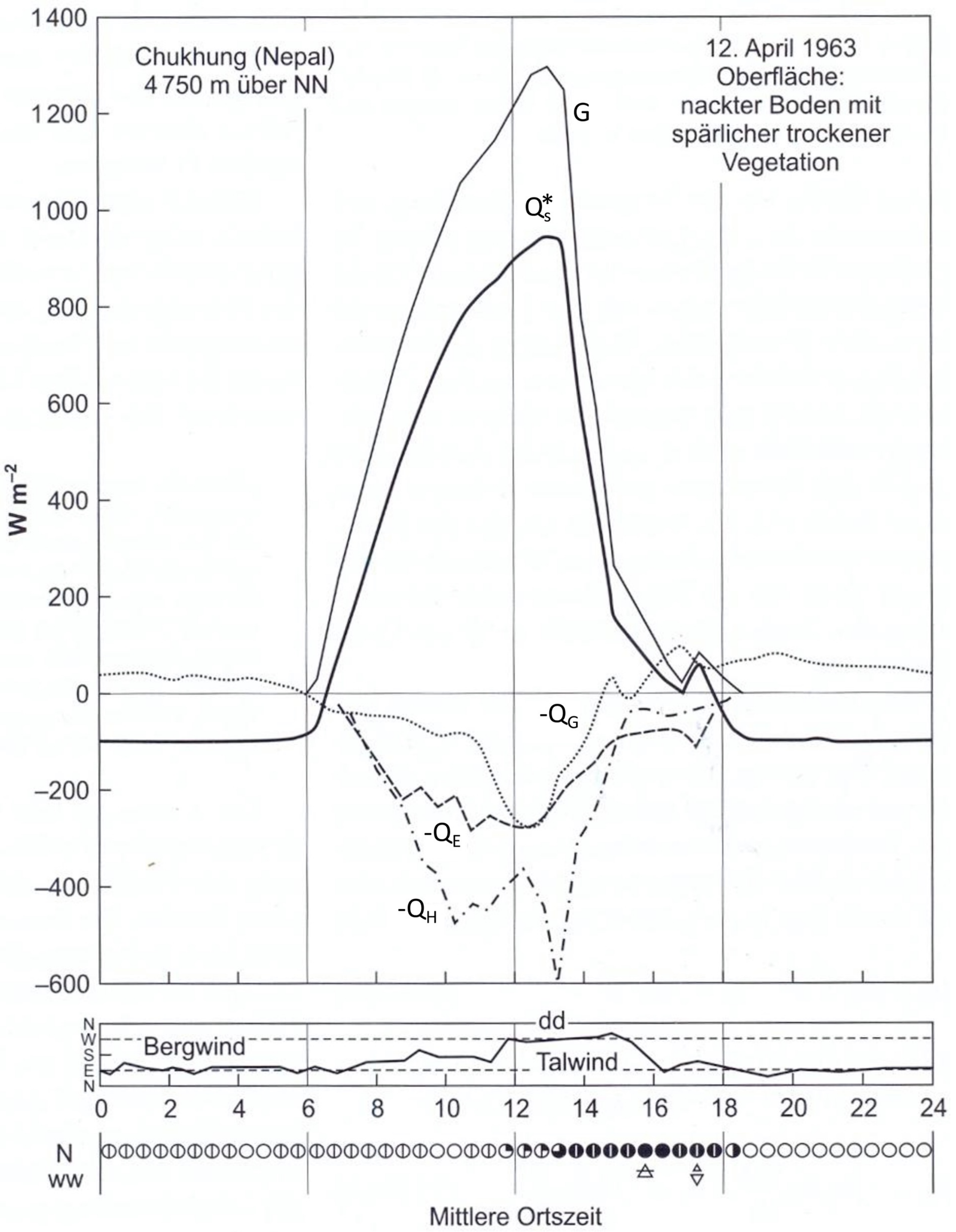
What are the relevant fluxes?

Energy Balance Terms



Vertical cross-section sketch of net radiative input to the surface Flux, and resulting heat fluxes (modified after Wallace and Hobbs, 2006).

Energy Balance Terms



Daily cycle of the energy balance components above moraine material in Nepal (modified after Krauss, 2008).

Fluxes				Situation
Q_S	Q_G	Q_H	Q_E	Generic case
Q_S	Q_G	Q_H		Dry soil
Q_S	Q_G		Q_E	Ocean, Lake
Q_S		Q_H	Q_E	Soil with low thermal conductivity
	Q_G	Q_H	Q_E	Night, cloud cover
Q_S	Q_G			Melting surface
Q_S		Q_H		Dry soil with low thermal conductivity
Q_S			Q_E	Intensive evapotranspiration
	Q_G	Q_H		Dry soil at night with cloud cover
	Q_G		Q_E	Ocean, Lake at cloudy night

Contribution of fluxes in different environments (Krauss, 2008)

	Q_S (Wm^{-2})	Q_G (Wm^{-2})	Q_H (Wm^{-2})	Q_E (Wm^{-2})	$W*$ $(mm\ d^{-1})$	Q_H/Q_E
Forest	154	6	77	72	2.5	1.07
Potatoes	121	3	27	91	3.1	0.30
Alfalfa	131	1	3	127	4.4	0.02

Surface flux measurements during daytime (Krauss, 2008)

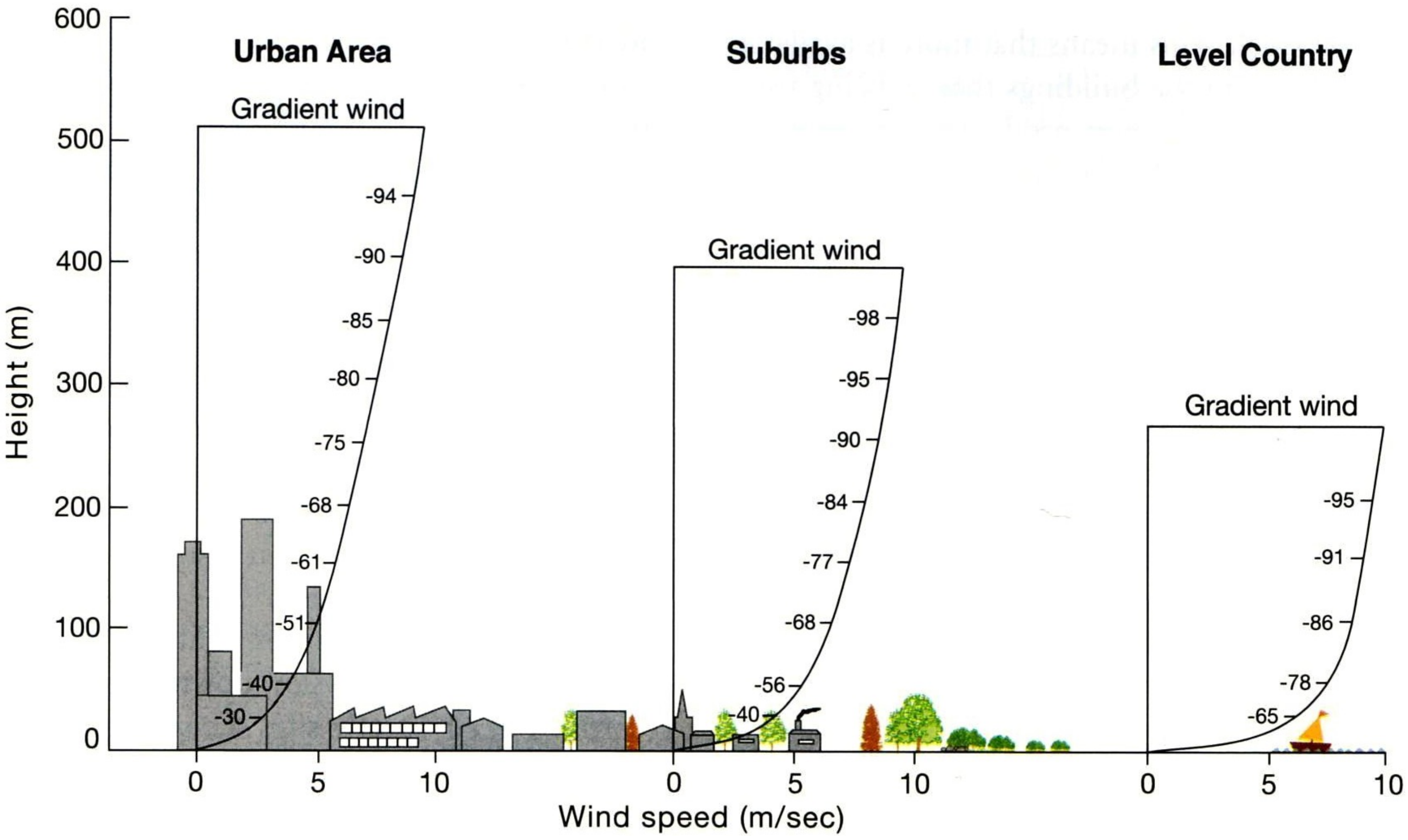
Reminder

Properties

- K-Theory for the parametrization of turbulent fluxes (surface layer)
- Mixing-Length theory
- Surface energy balance
- Parametrization of the radiation
- Parametrization of the surface fluxes (Bulk approach, Bowen-Ratio)
- Bulk coefficients, roughness length
- Ground heat flux

Logarithmic Wind Profile

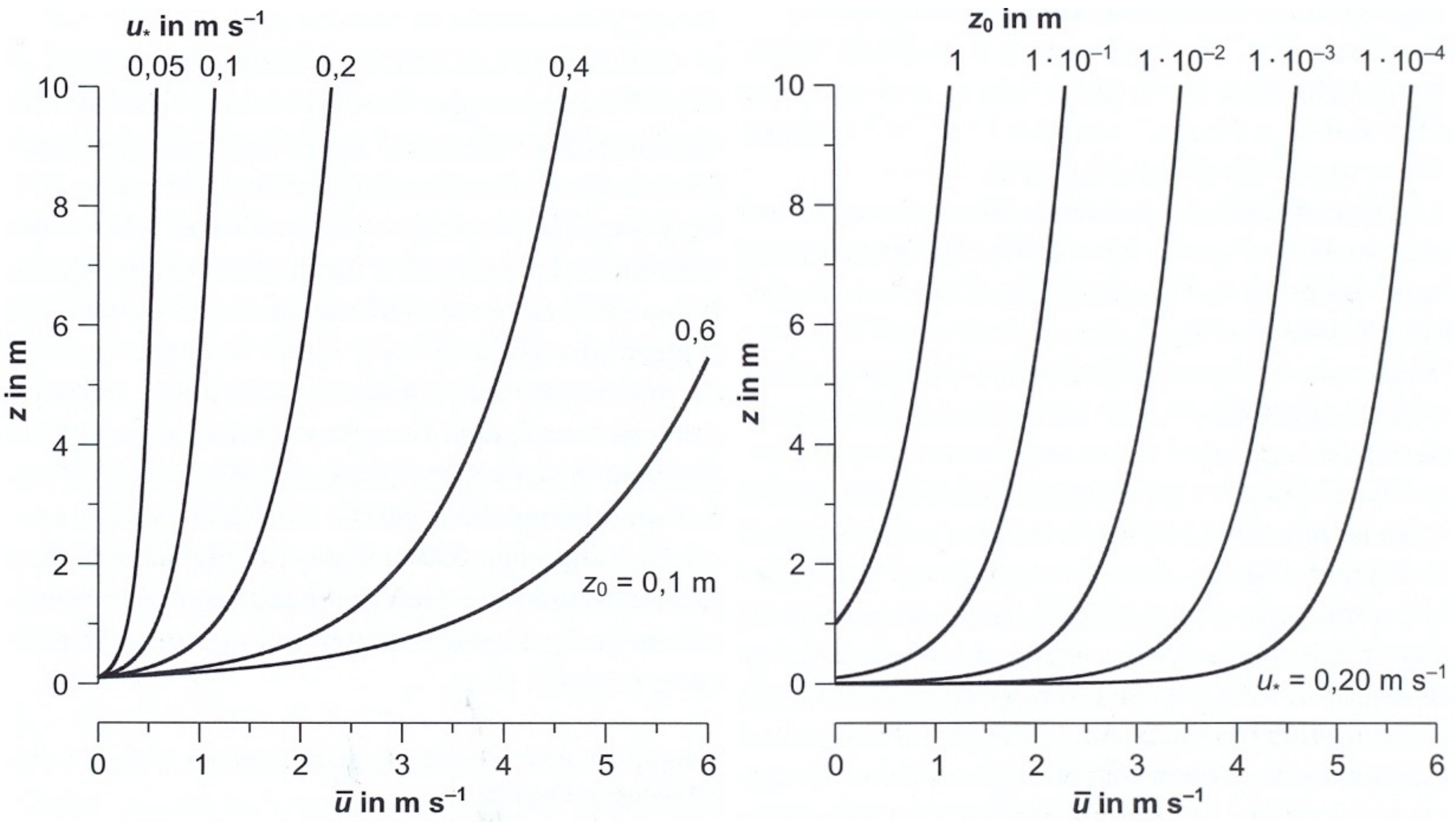
Logarithmic wind profile - Neutral case



$$u(z) = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right)$$

with u_* : friction velocity [m s^{-1}]
κ : Von Karman constant (0.4)
z_0 : Roughness length [m]

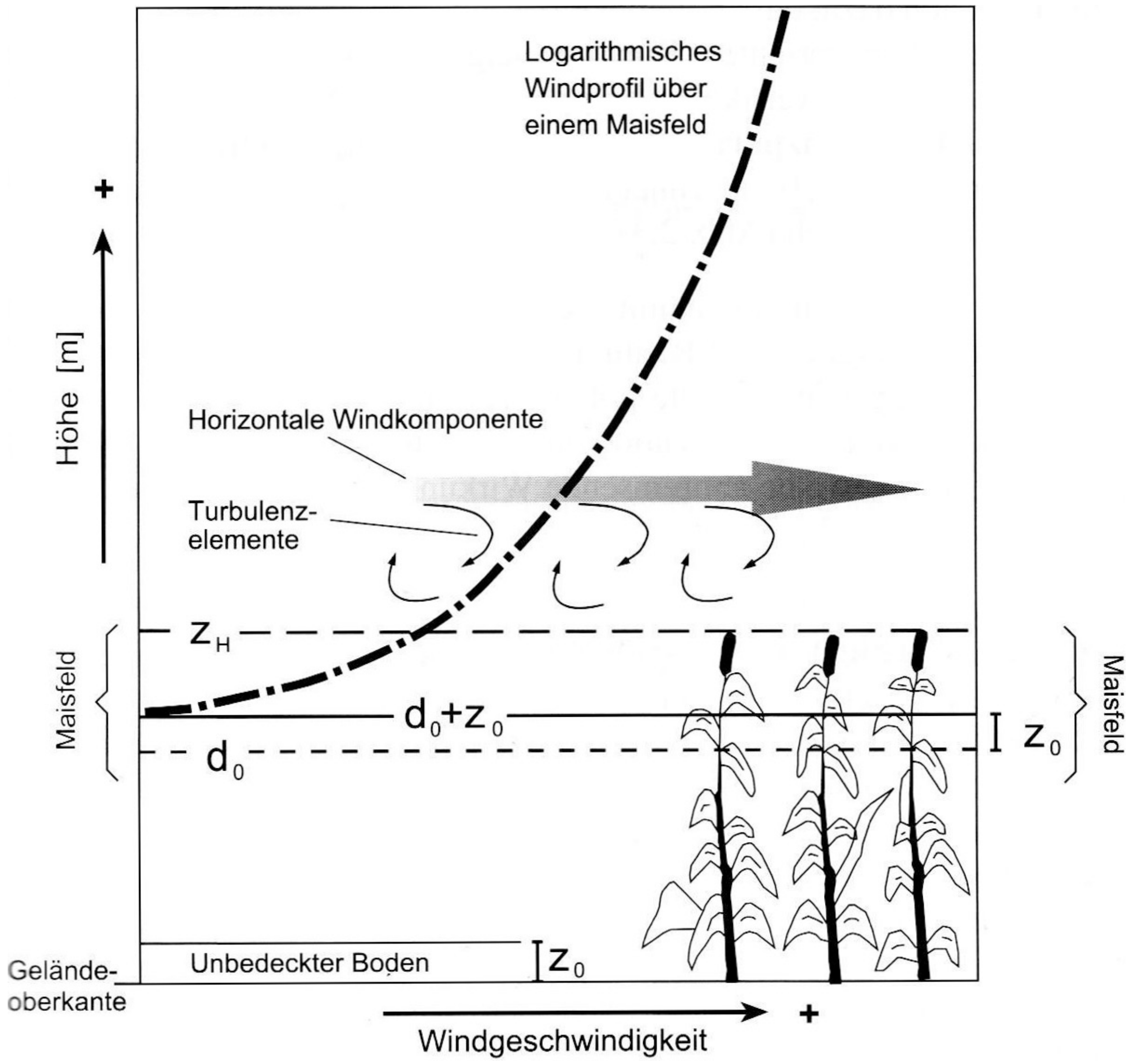
Logarithmic wind profile - Neutral case



$$u(z) = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right)$$

Wind profiles for neutral stability with constant roughness length (left) and friction velocity (right).

Logarithmic wind profile – Displacement height



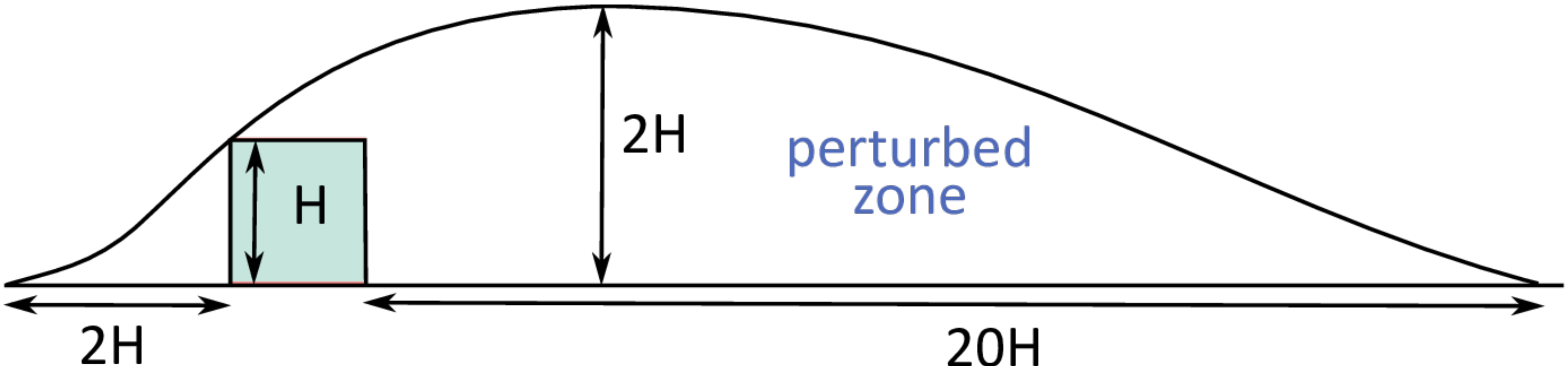
$$u(z) = \frac{u_*}{\kappa} \ln\left(\frac{z - d}{z_0}\right)$$

Influence of the surface structure on the vertical wind profile (Bendix, 2004)

The logarithmic law applied only for

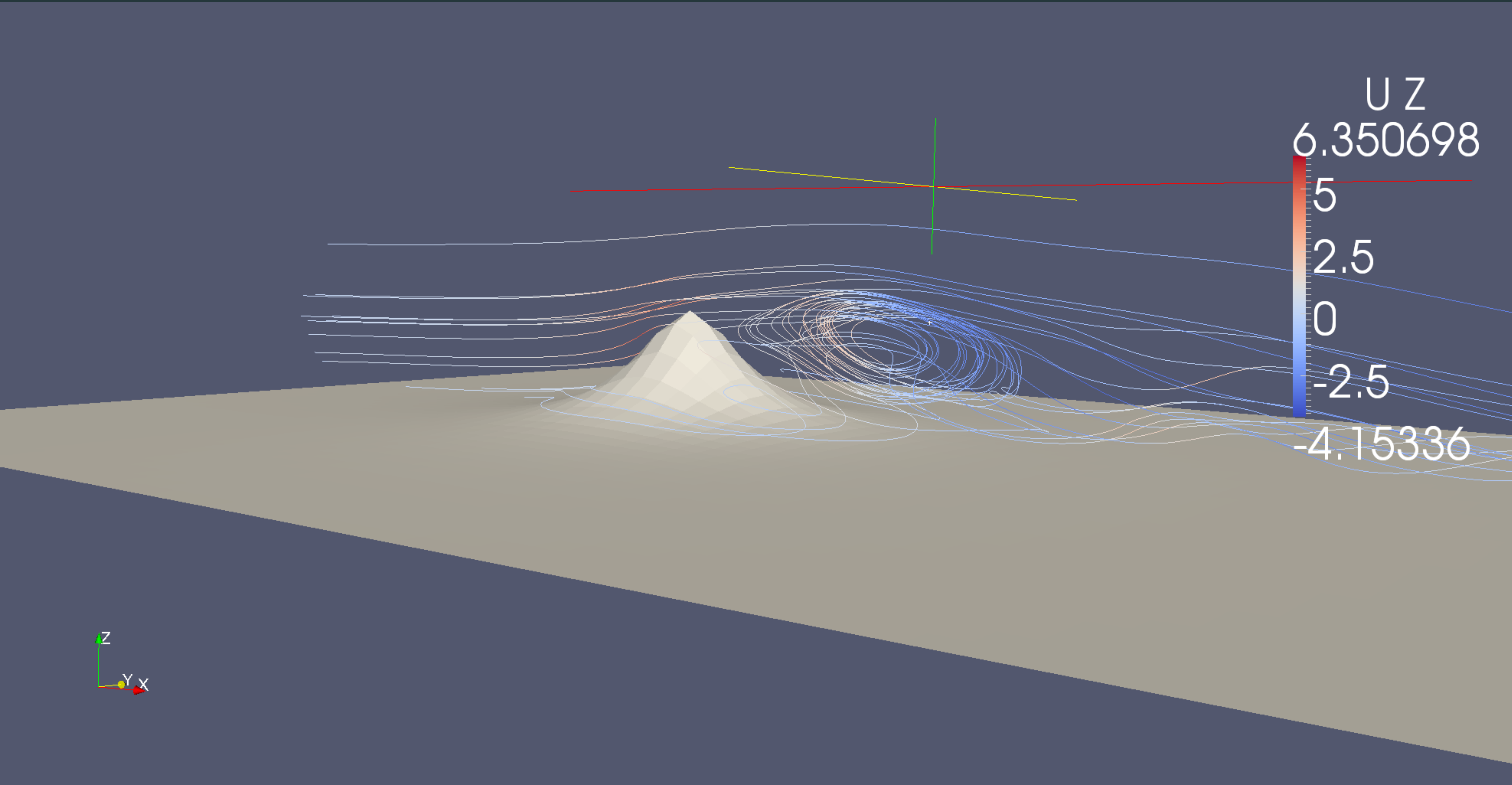
- homogeneous and stationary flow
- neutral stable atmospheric conditions
- height $z > 20 \cdot z_0 + d$
- locations far from obstacles, i.e. $> 20 \cdot H$ from wake, and $> 2 \cdot H$ upstream from obstacle

Obstacles

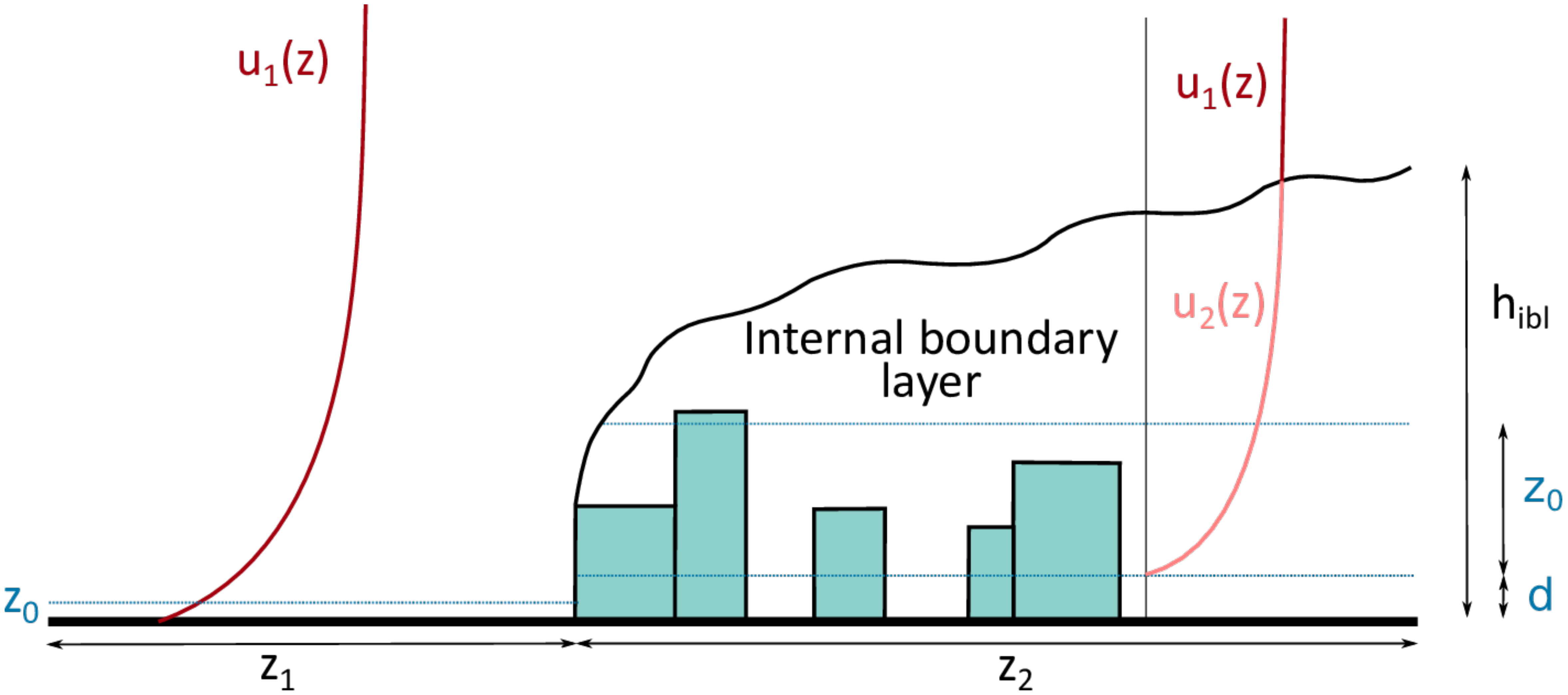


The effect of terrain roughness on the wind speed profile.

Obstacles



Internal boundary layer



Development of an internal boundary layer at the land-city transition.

Practical relevance of the IBL

- Profile measurements above the boundary layer
- Wind shield for soil erosion
- Wind energy

Reminder

- Logarithmic wind profile
- Displacement height and roughness length
- Boundary condition
- Internal boundary layer