Short Summer School on Atmospheric Physics

Surface -Atmosphere Interactions

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Summary

- Surface energy budget
- Surface mass budget
- Soil heat and water transfer
- Vegetation
- Turbulent fluxes near surface
- Eddy covariance measurements
- Snow and Water surfaces

Note

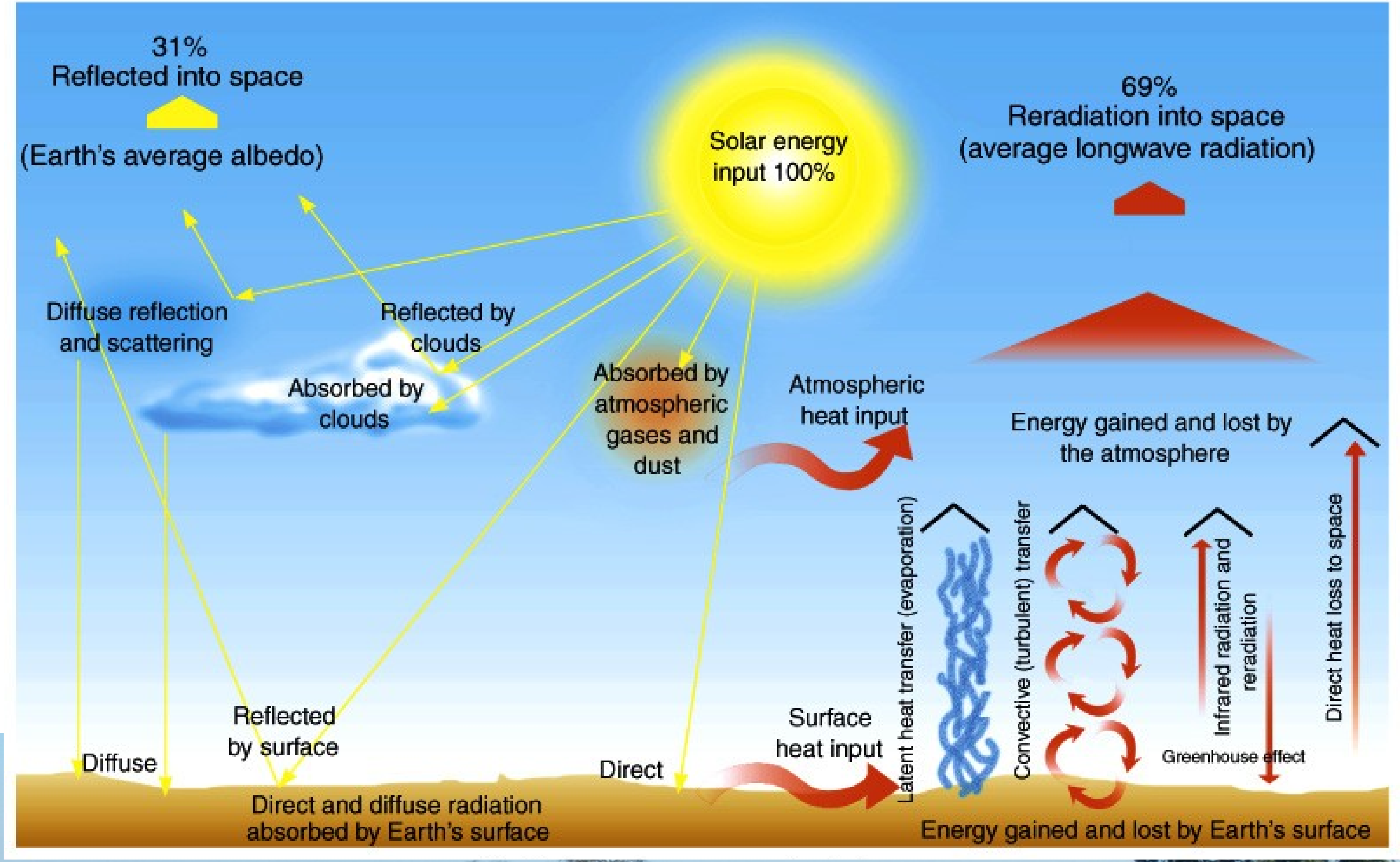
- Some of the slides included in this presentation were tacked from:
  - ECMWF NWP training material 2015:
  - Land Surface (1): Introduction from Gianpaolo Balsamo
  - Land Surface (2): Surface Energy, Water Cycle from Gianpaolo Balsamo
  - Land Surface (3):Snow from Emanuel Dutra
  - "A Brief Practical Guide to Eddy Covariance Flux Measurements: Principles and Work flow Examples for Scientific and Industrial Applications"

  by G. Burba and D. Anderson of LI-COR Biosciences

Earth energy cascade

- The sun emits 4 x 10^{26} W
- the Earth intercepts 1.37 kW/m^2
- This energy is distributed between
  - Direct reflection (~30%)
  - Conversion to heat, mostly by surface absorption (~43%), re-radiated in the infrared
  - Evaporation, Precipitation, Runoff (~22%)
  - Rest of the processes (~5%, Winds, Waves, Convection, Currents, Photosynthesis, Organic decay, tides, …)

Global Energy Budget



Global Energy budget

- Surface fluxes and the atmosphere
  - Sensible heat (*H*) at the bottom means energy immediately available close to the surface
  - Latent heat (*LE*) means delayed availability through condensation processes, for the whole tropospheric column
  - The net radiative cooling of the whole atmosphere is balanced by condensation and the sensible heat flux at the surface. Land surface processes affect directly (*H*) or indirectly (condensation, radiative cooling, …) this balance.

Energy Budget

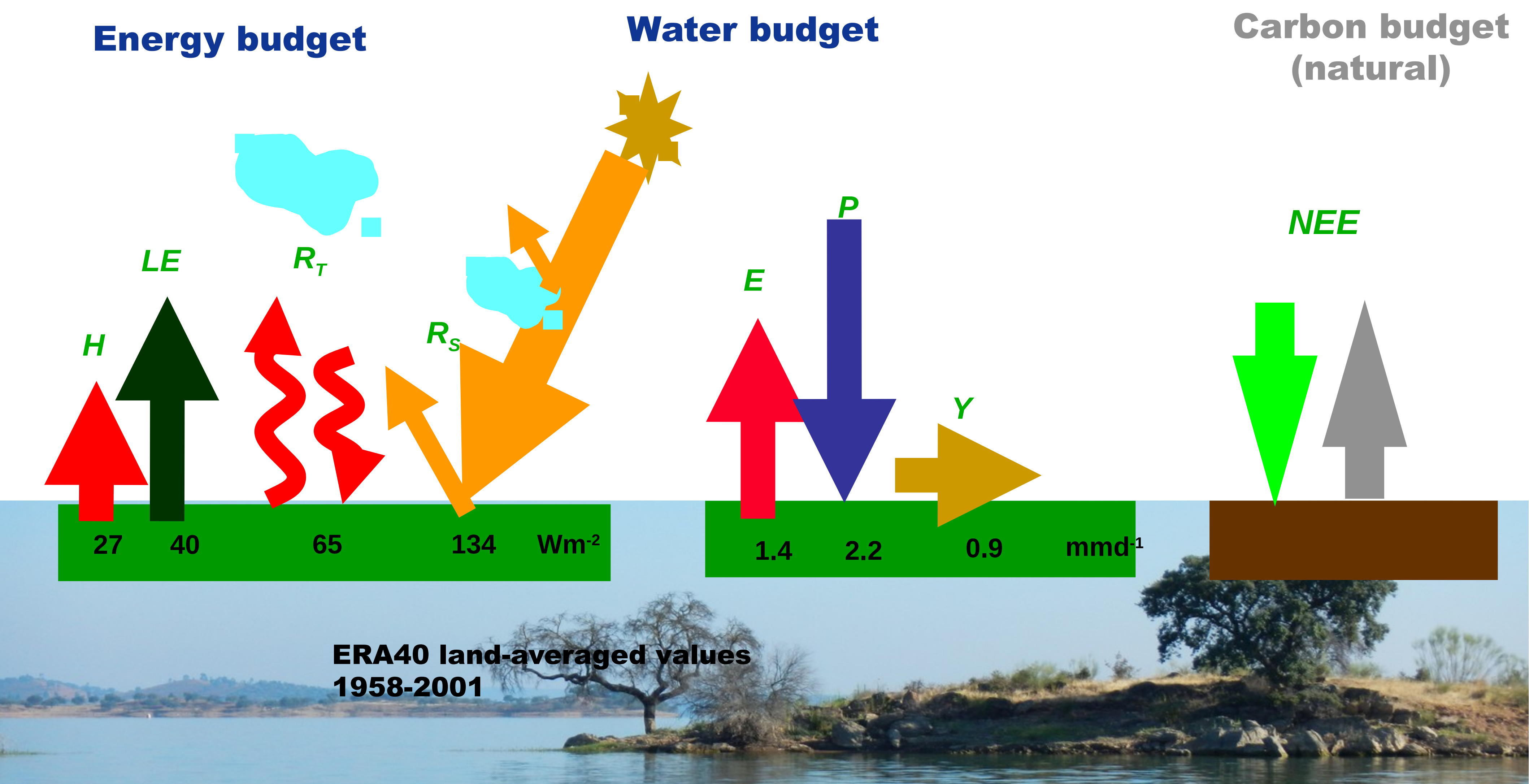
- Mean surface energy fluxes (Wm^{-2}) in the ERA40 atmospheric reanalysis (1958-2001); positive fluxes downward

	R_S	R_T	H	LE	G	$Bo=H/LE$
Land	134	-65	-27	-40	2	0.7
Sea	166	-50	-12	-102	3	0.1

- Land surface
  - The net radiative flux at the surface (R_S+R_T) is downward. Small storage at the surface (G) implies upward sensible and latent heat fluxes.
- Bowen ratio: Land vs Sea
  - Different physical mechanisms controlling the exchanges at the surface
    - Continents: Fast responsive surface; Surface temperature adjusts quickly to maintain zero ground heat flux
    - Oceans: Large thermal inertia; Small variations of surface temperature allowing imbalances on a much longer time scale

Role of land surface

- Atmospheric general circulation models need boundary conditions for the enthalpy, moisture (and momentum) equations: Fluxes of energy, water at the surface.



Conservation of Energy at the Surface

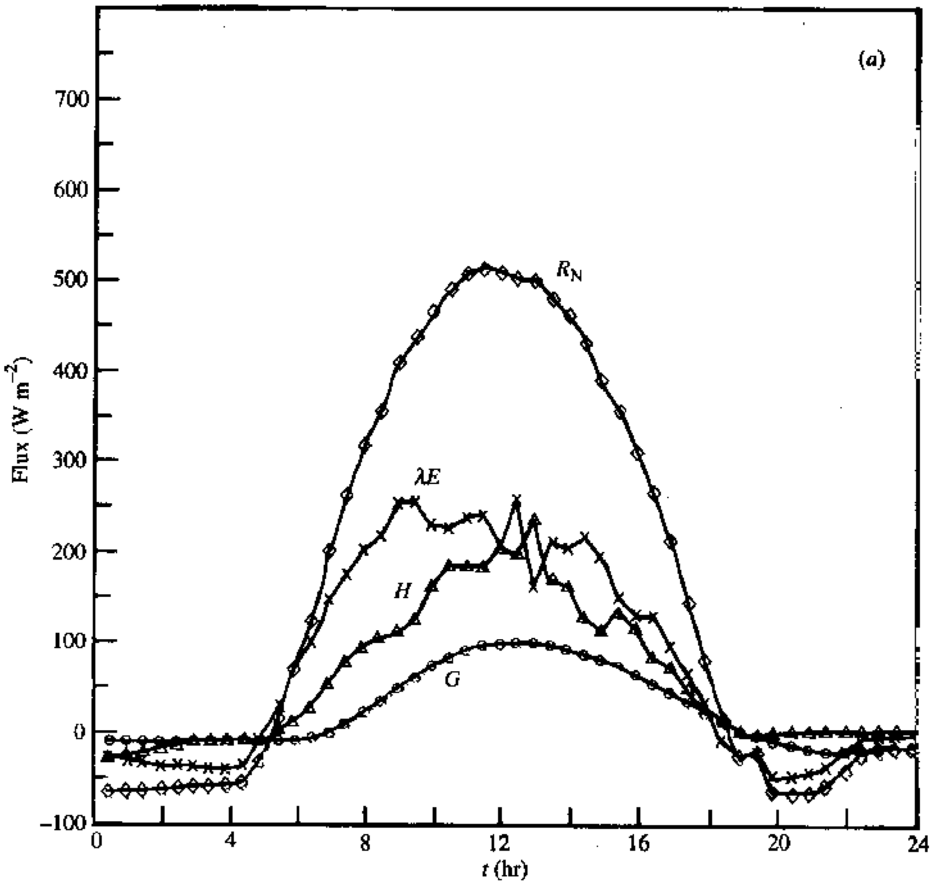
$$C\frac{\partial T_S}{\partial t}=R_N-H-LE-G$$ Prognostic Equation

H is the Sensible Heat Flux,
LE is the Latent Heat Flux,
G is the Ground Heat Flux,
R_N is the net radiative Flux,

$C\frac{\partial T_S}{\partial t}$ is the accumulated energy at a thin soil layer (per unit area and time).
C is a surface heat capacity
(C = $C_T\delta z$, where C_T is the volumetric heat capacity

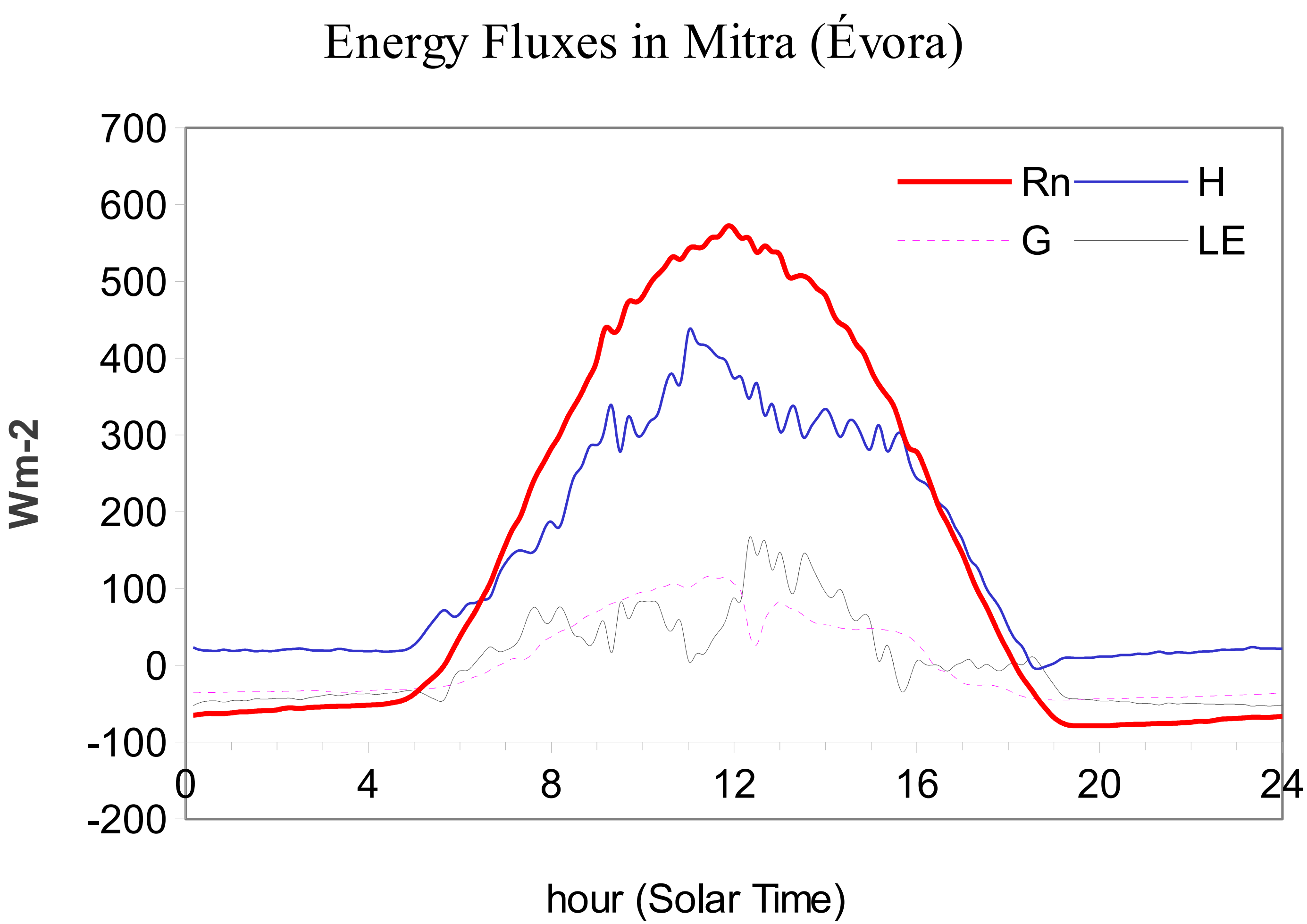
$$R_N-H-LE-G=0$$ Diagnostic Equation

Energy budget: Example

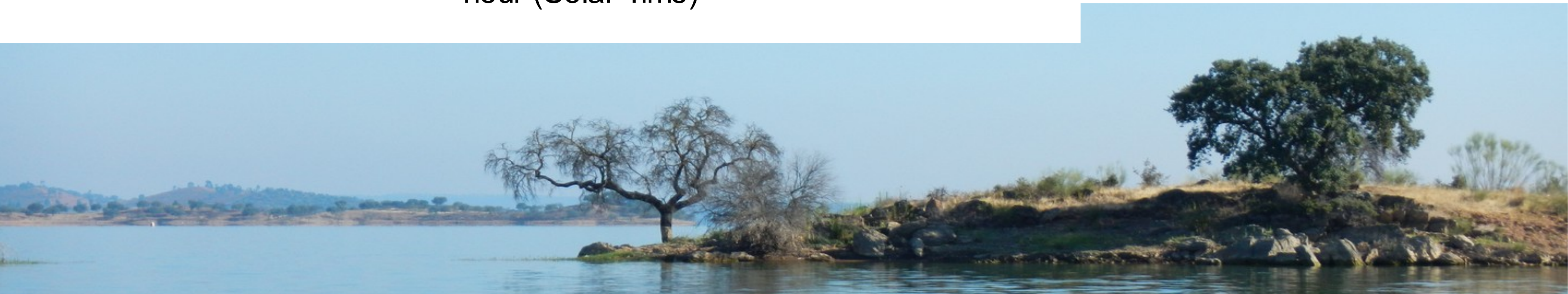


- Observed energy budget over a maize field on a sunny summer day in France
- From Noilhan and Planton, 1989

Energy budget: Example (2)



- Mean daily cycles of energy fluxes observed in Mitra (Évora) in July1994



$$R_N = (1-\alpha)R_G + \varepsilon(R_A - \sigma T_s^4)$$

Global solar radiation and the atmospheric radiation are inputs (boundary condition) to compute the energy conservation equation at the surface.

- α is the albedo and ε is the emissivity. These parameters depends:
  - Soil type
  - vegetation cover
  - soil water content
  - zenith angle

The surface radiation

Surface albedo

Surface emissivity

Skin temperature

- In some cases (snow, sea ice, dense canopies) the impinging solar radiations penetrates the “ground” layer and is absorbed at a variable depth. In those cases, an extinction coefficient is needed.

Table 3.1

Radiative Properties of Natural Surfaces[a]

Surface type	Other specifications	Albedo (a)	Emissivity (ε)
Water	Small zenith angle	0.03–0.10	0.92–0.97
	Large zenith angle	0.10–0.50	0.92–0.97
Snow	Old	0.40–0.70	0.82–0.89
	Fresh	0.45–0.95	0.90–0.99
Ice	Sea	0.30–0.40	0.92–0.97
	Glacier	0.20–0.40	
Bare sand	Dry	0.35–0.45	0.84–0.90
	Wet	0.20–0.30	0.91–0.95
Bare soil	Dry clay	0.20–0.35	0.95
	Moist clay	0.10–0.20	0.97
	Wet fallow field	0.05–0.07	
Paved	Concrete	0.17–0.27	0.71–0.88
	Black gravel road	0.05–0.10	0.88–0.95
Grass	Long (1 m) Short (0.02 m)	0.16–0.26	0.90–0.95
Agricultural	Wheat, rice, etc.	0.10–0.25	0.90–0.99
	Orchards	0.15–0.20	0.90–0.95
Forests	Deciduous	0.10–0.20	0.97–0.98
	Coniferous	0.05–0.15	0.97–0.99

[a] Compiled from Sellers (1965), Kondratyev (1969), and Oke (1978).

Arya, 1988

Soil heat transfer

In the absence of phase changes, heat conduction in the soil obeys a Fourier law

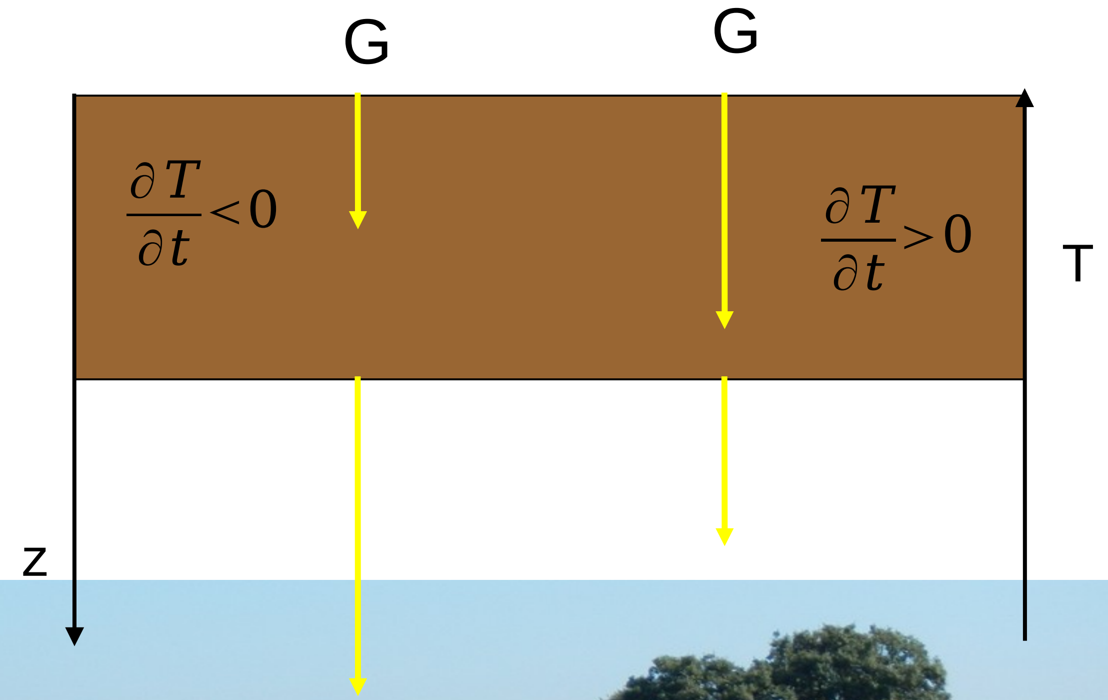
$$C_T \frac{\partial T}{\partial t} = \frac{-\partial G}{\partial z} = \frac{\partial}{\partial z} K_T \frac{\partial T}{\partial z}$$

G Soil heat flux

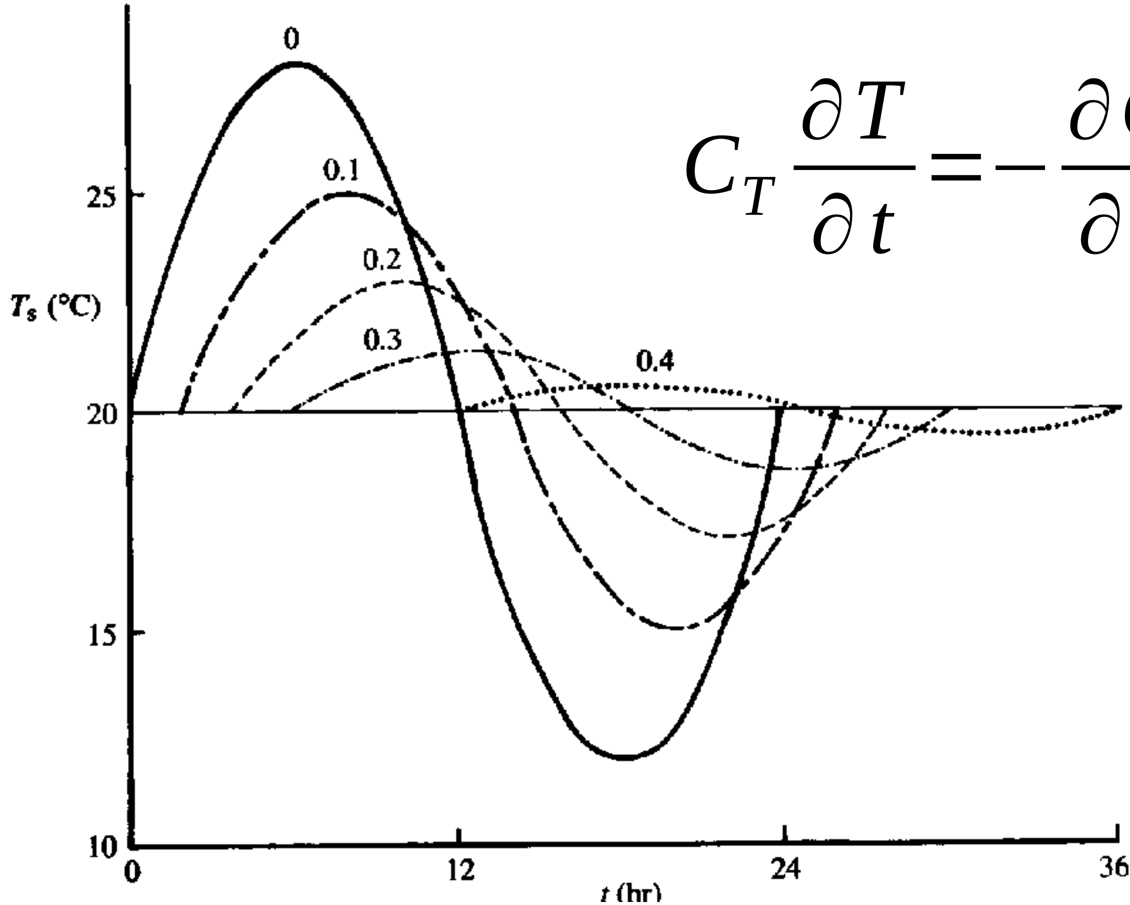
C_T Soil volumetric heat capacity

K_T Thermal conductivity

These parameters depends on the soil texture and water content!



Heat Flux in the soil

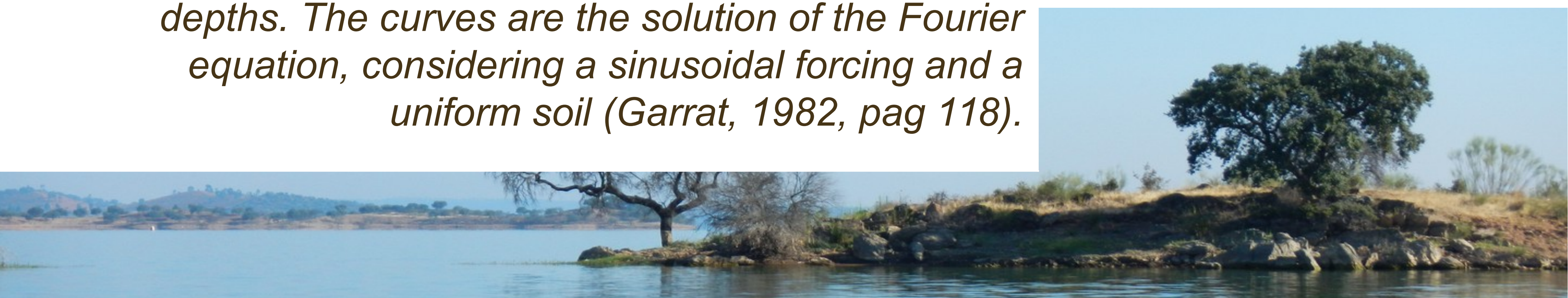


$$C_T \frac{\partial T}{\partial t} = -\frac{\partial G}{\partial z} = \frac{\partial}{\partial z} K_T \frac{\partial T}{\partial z}$$

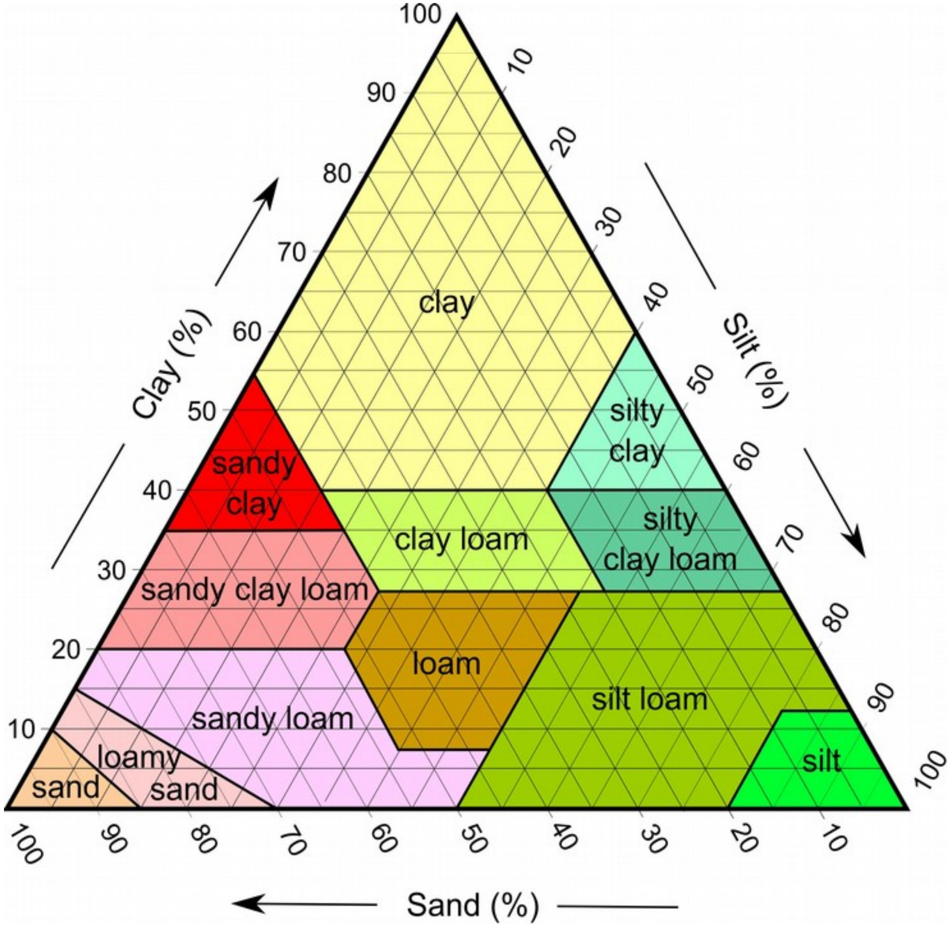
K_T = Cte in z

$$C_T \frac{\partial T}{\partial t} = K_T \frac{\partial^2 T}{\partial z^2}$$

Figure: Daily Cycle of soil temperature at different depths. The curves are the solution of the Fourier equation, considering a sinusoidal forcing and a uniform soil (Garrat, 1982, pag 118).



Soil temperature and soil texture



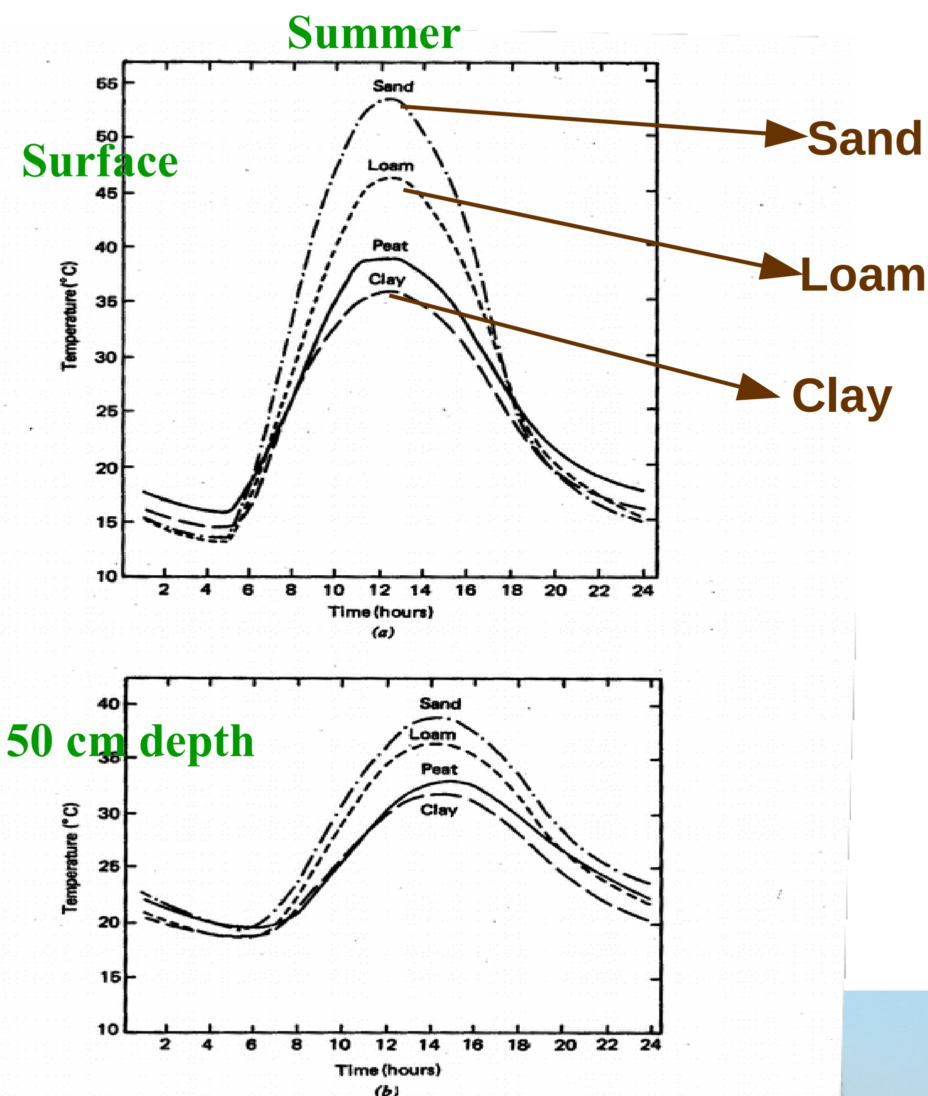


Fig. 2.6 Daily course of temperature (*a*) at the surface and (*b*) at a depth of 50 mm on clear summer days at Sapporo, Japan (after Yakuwa, 1946).

Rosenberg et al 1983

Surface Water Budget

$$\frac{\partial W}{\partial t} = P - R - E - F_W$$

- $\frac{\partial W}{\partial t}$: Change in water content in the surface layer
- P: Precipitation
- R: Runoff
- E: Evapotranspiration
- Soil water flux

Soil Water transfer

The Thermal proprieties depends also on the soil water content, so it is necessary to compute simultaneously the soil water difusion.

Hillel 1982

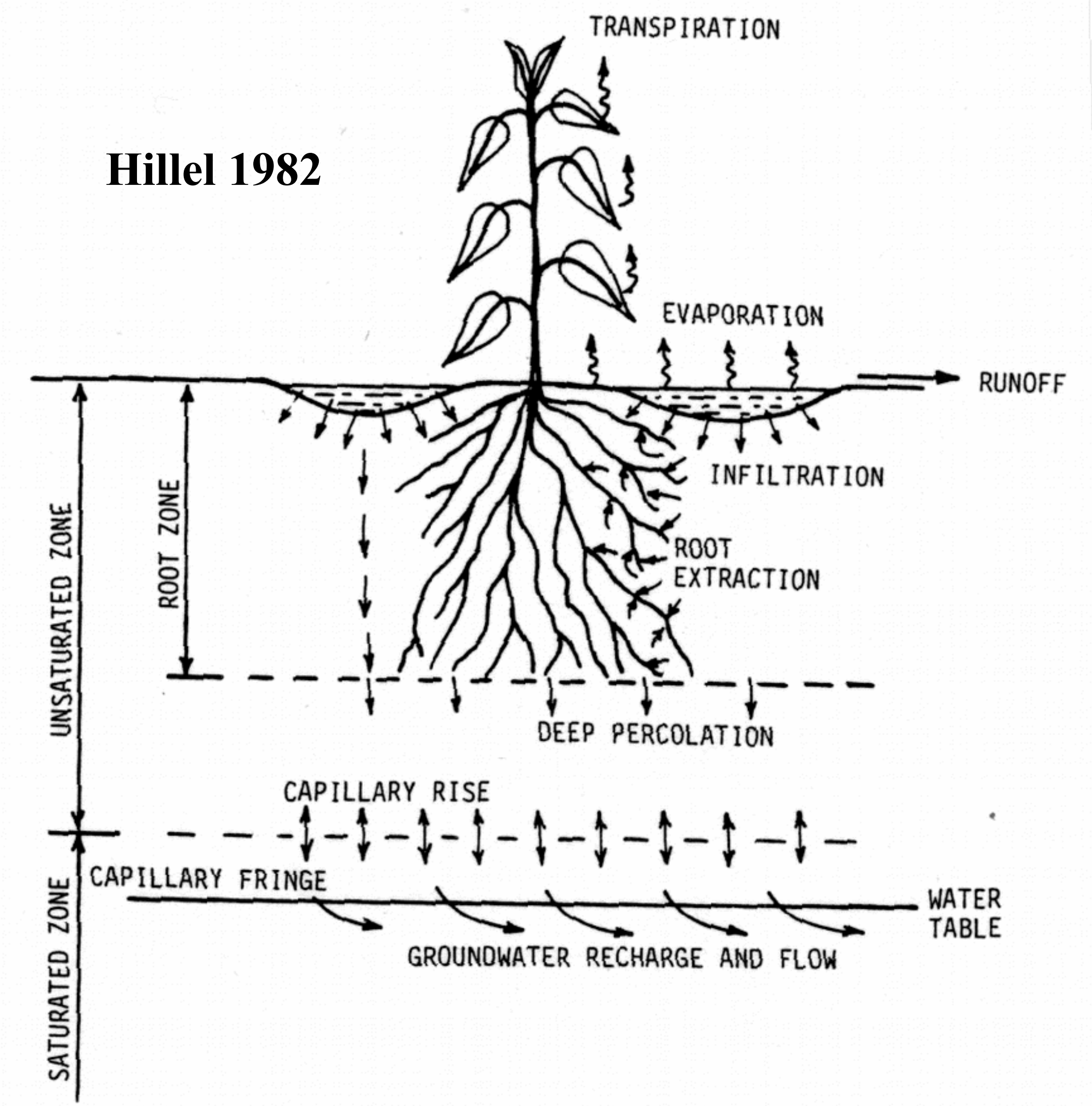


Fig. 17.1. The water balance of a root zone (schematic).

$$\rho_W \frac{\partial w}{\partial t} = \frac{\partial F_w}{\partial z} + \rho_W S_W$$

- w soil water content ($m^3 m^{-3}$)
- F_w Soil water Flux ($kg\, m^{-2} s^{-1}$)
- Soil water source/sink, ie: Root extraction – The amount of water transported from the root system up to the stomata (due to the difference in the osmotic pressure) and then available for transpiration

Soil water properties

Hillel 1982

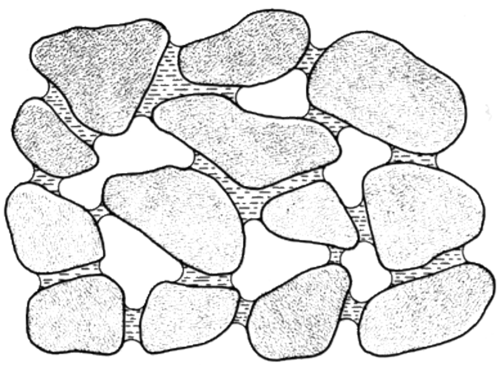


Fig. 7.1. Water in an unsaturated coarse-textured soil.

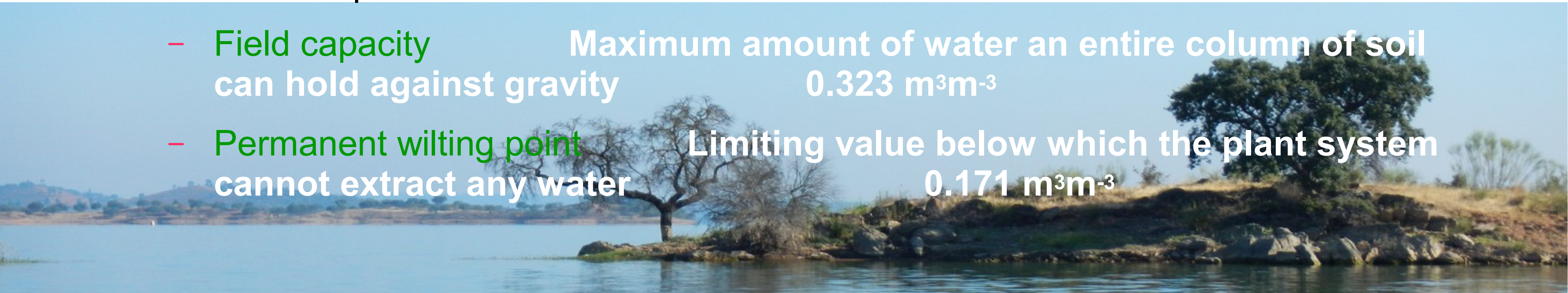
TABLE I

Jacquemin and Noilhan 1990

Critical water contents of soils derived from the classification of Clapp and Hornberger (1978): saturated moisture w_{sat}, field capacity w_{fl}, wilting point w_{wilt}. The field capacity is associated with a hydric conductivity of 0.1 mm/day. The wilting point corresponds to a moisture potential of −15 bar

Soil type	w_{sat} (m^3/m^3)	w_{fc} (m^3/m^3)	w_{wilt} (m^3/m^3)
Sand	0.395	0.135	0.068
Loamy sand	0.410	0.150	0.075
Sandy loam	0.435	0.195	0.114
Silt loam	0.485	0.255	0.179
Loam	0.451	0.240	0.155
Sandy clay loam	0.420	0.255	0.175
Silty clay loam	0.477	0.322	0.218
Clay loam	0.476	0.325	0.250
Sandy clay	0.426	0.310	0.219
Silty clay	0.482	0.370	0.283
Clay	0.482	0.367	0.286

- 3 numbers defining soil water properties
  - Saturation (soil porosity) Maximum amount of water that the soil can hold when all pores are filled 0.472 m^3m^{-3}
  - Field capacity Maximum amount of water an entire column of soil can hold against gravity 0.323 m^3m^{-3}
  - Permanent wilting point Limiting value below which the plant system cannot extract any water 0.171 m^3m^{-3}



Turbulent Heat and Water Fluxes

Recap:

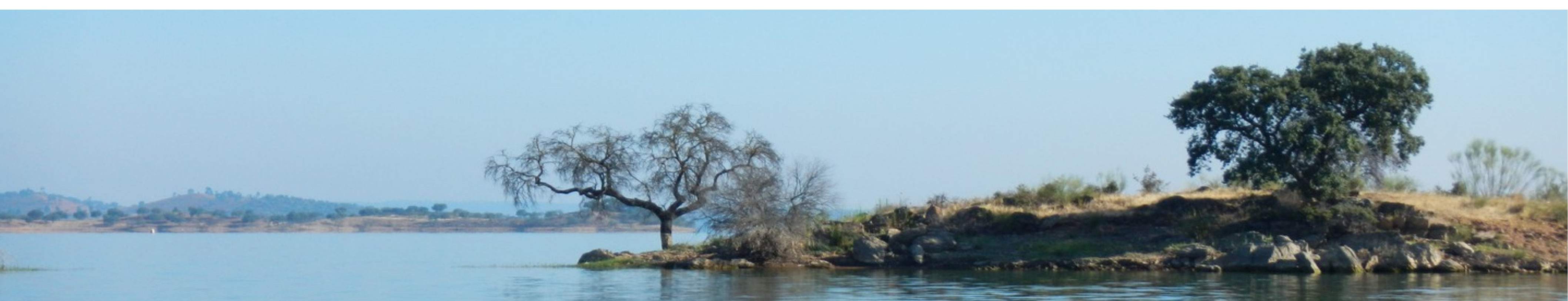
$$C\frac{\partial T_S}{\partial t}=R_N-H-LE-G$$

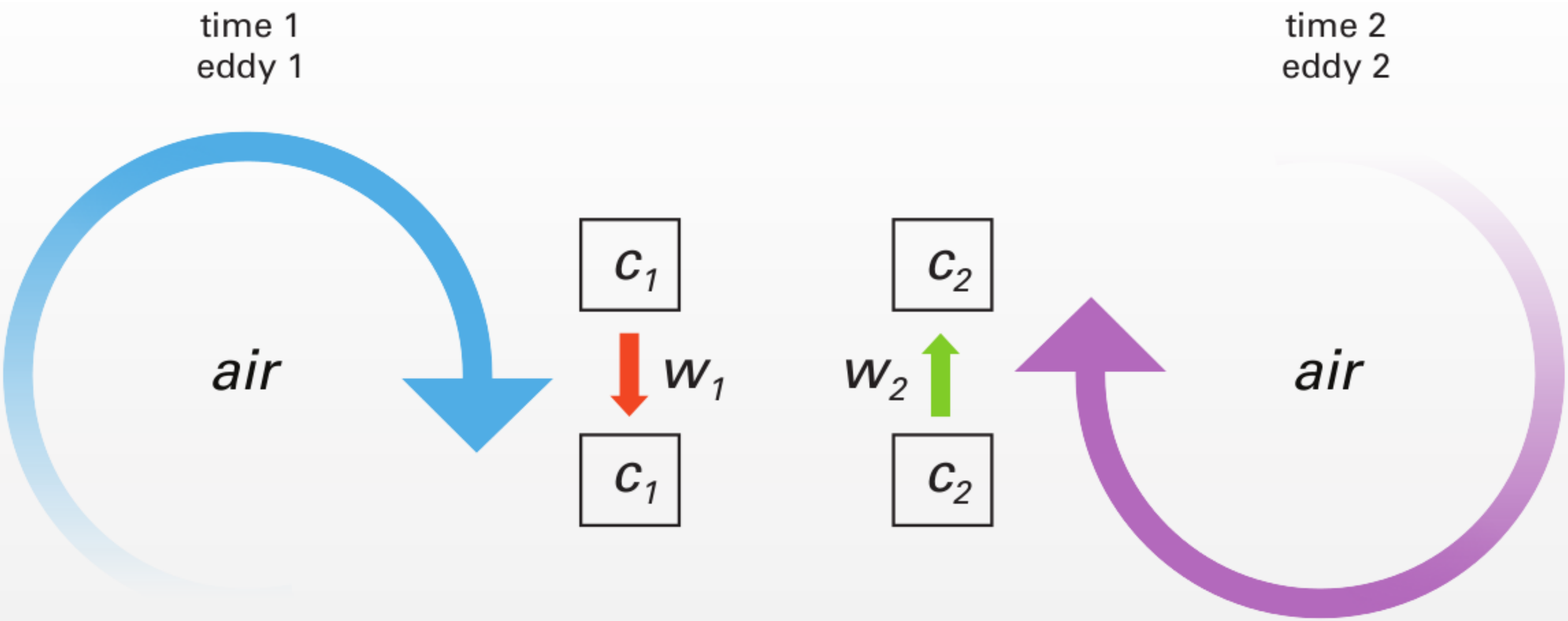
$$\frac{\partial W}{\partial t}=P-R-E-F_W$$

H and E are turbulent Fluxes !
We will se how to compute it

Eddys

- Air flow can be imagined as a horizontal flow of numerous rotating eddies
- Each eddy has 3-D components, including a vertical wind component
- The diagram looks chaotic, but components can be measured from a tower





At a single point on the tower:

Eddy 1 moves parcel of air c_1 down with the speed w_1, then eddy 2 moves parcel c_2 up with the speed w_2

Each parcel has concentration, temperature, humidity; if we know these and the speed – we know the flux

Principles

The physical principle:

If we know how many molecules went up with eddies at time 1, and how many molecules went down with eddies at time 2 at the same point – we can calculate vertical flux at that point and over that time period

The mathematical principle:

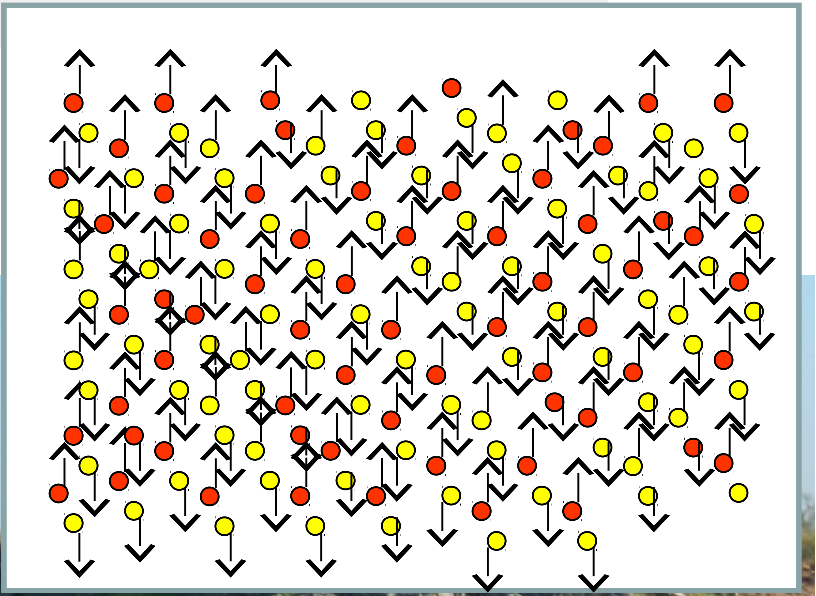
Vertical flux can be represented as a covariance of the vertical velocity and concentration of the entity of interest

The instrument challenge:

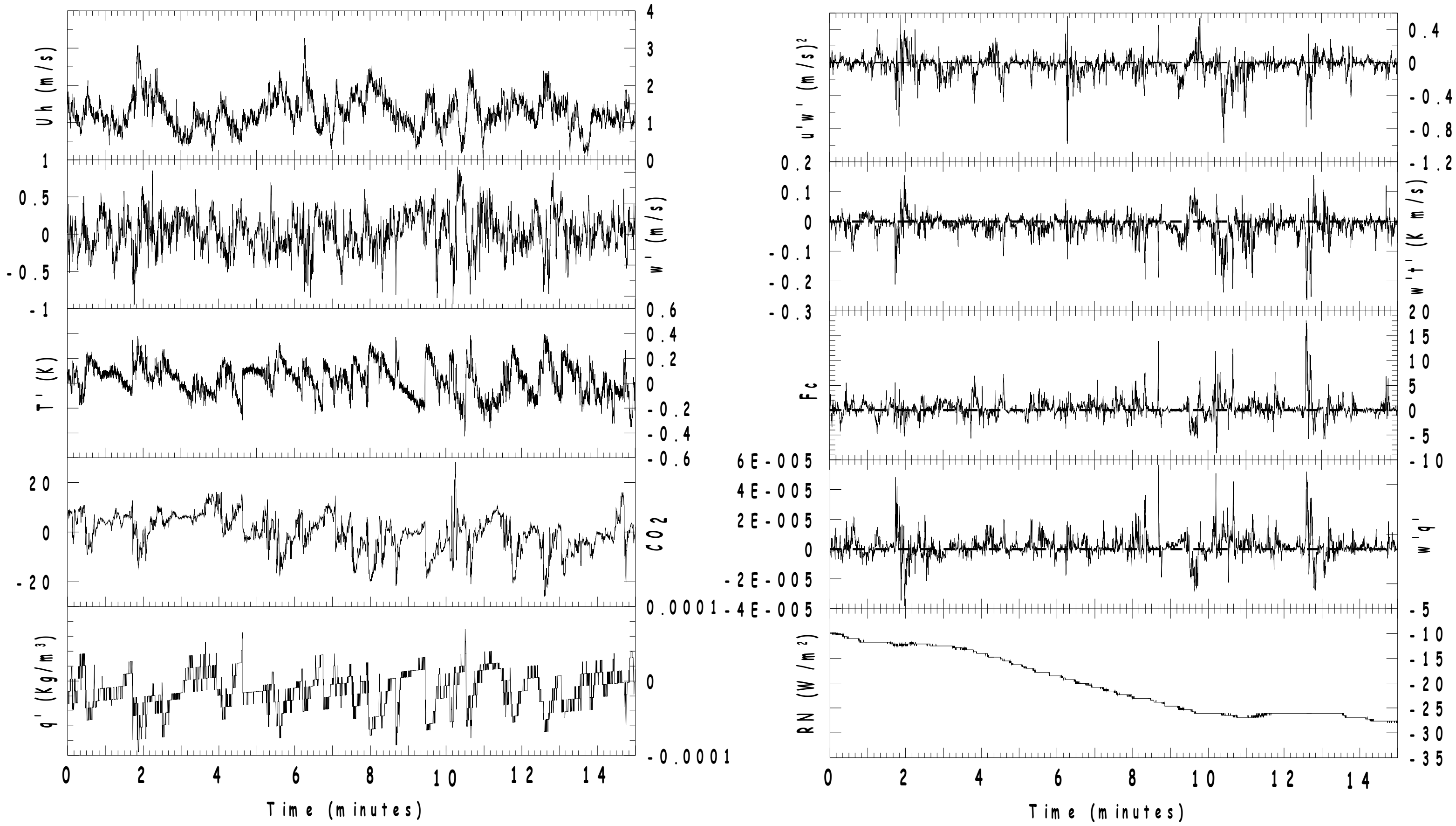
Turbulent fluctuations occur very rapidly, so measurements of up-and-down movements and of the number of molecules should be done very rapidly

$$\overline{\theta' w'}$$

$$\overline{u' w'}$$



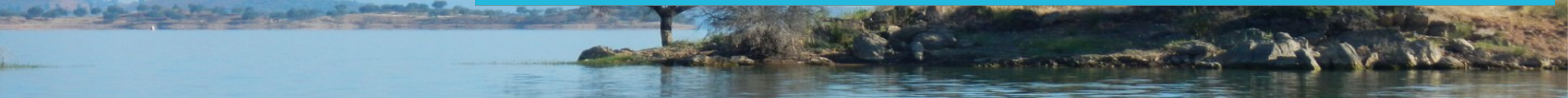
Turbulence



Reynolds decomposition:

$$u = \overline{u} + u' \quad v = \overline{v} + v' \quad w = \overline{w} + w'$$

$$\theta = \overline{\theta} + \theta' \quad q = \overline{q} + q' \quad c = \overline{c} + c'$$



In turbulent flow, vertical flux can be presented as:
(s is the dry mole fraction of the gas of interest in the air)

$$F = \overline{\rho_d w s}$$

Reynolds decomposition is then used to break terms into means and deviations:

$$F = \overline{(\bar{\rho}_d + \rho'_d)(\bar{w} + w')(\bar{s} + s')}$$

Opening the parentheses:

$$F = \overline{(\bar{\rho}_d \bar{w}\bar{s} + \cancel{\bar{\rho}_d \bar{w} s'} + \cancel{\bar{\rho}_d w' \bar{s}} + \bar{\rho}_d w' s' + \cancel{\rho'_d \bar{w}\bar{s}} + \rho'_d \bar{w} s' + \rho'_d w' \bar{s} + \rho'_d w' s')}$$

averaged deviation from the average is zero

Equation is simplified:

$$F = (\bar{\rho}_d \bar{w}\bar{s} + \bar{\rho}_d \overline{w's'} + \bar{w}\,\overline{\rho'_d s'} + \bar{s}\,\overline{\rho'_d w'} + \overline{\rho'_d w' s'})$$

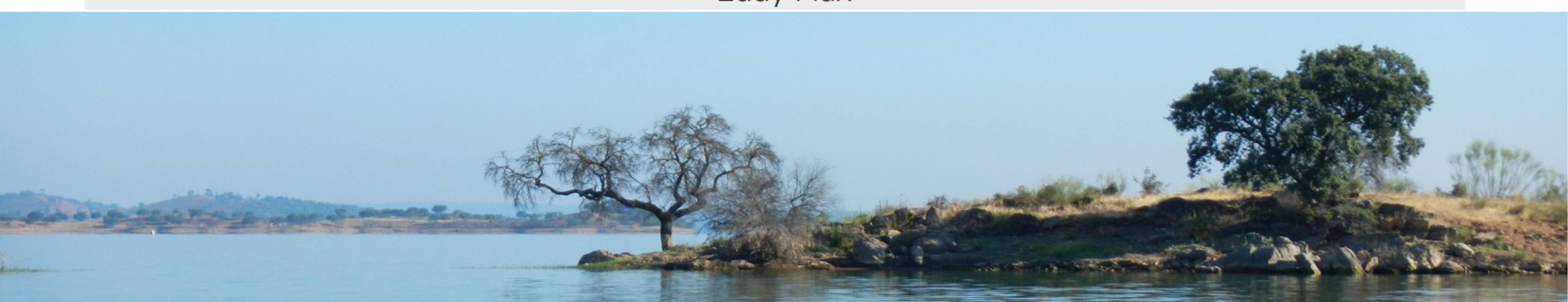
Now an important assumption is made (for conventional eddy covariance) – air density fluctuations are assumed to be negligible:

$$F = (\overline{\rho}_d \overline{w}\,\overline{s} + \overline{\rho}_d \overline{w's'} + \cancel{\overline{w}\,\overline{\rho_d' s'}} + \cancel{\overline{s}\,\overline{\rho_d' w'}} + \cancel{\overline{\rho_d' w' s'}}) = \cancel{\overline{\rho}_d \overline{w}\,\overline{s}} + \overline{\rho}_d \overline{w's'}$$

Then another important assumption is made – mean vertical flow is assumed to be negligible for horizontal homogeneous terrain (no divergence/convergence):

$$\boxed{F \approx \overline{\rho}_d \overline{w's'}}$$

'Eddy Flux'



Turbulent fluxes

momentum Flux (drag) $\tau = \rho \, \overline{u' w'}$

Sensible heat Flux $H = \rho \, c_p \, \overline{\theta' w'}$

Water vapor Flux (evaporation) $E = \rho \, \overline{q' w'}$

Any gas Flux $M = \overline{c' w'}$

Latent heat Flux $LE = \lambda E = \lambda \rho \, \overline{q' w'}$

Any gas transport is given by the mean flow transport + the turbulent transport $\overline{c w} = \bar{c} \, \overline{w} + \overline{c' w'}$

Parameterization of turbulent Fluxes

First order parameterization

$$\tau = \rho\, \overline{u' w'} = -\rho K_m \frac{\partial u}{\partial z}$$

analogy with molecular diffusion

K_m is the turbulent transfer coefficient for momentum (turbulent viscosity)

K_m is not a constant

Similarly:

$$\overline{\theta' w'} = -K_h \frac{\partial \theta}{\partial z} \qquad \overline{q' w'} = -K_q \frac{\partial q}{\partial z}$$

$K_{h,q}$ are turbulent transfer coefficients for sensible heat and for water

Parameterization of turbulent Fluxes

$$\tau = \rho C_D U^2$$

$$H = \rho c_p C_H U \left(\theta - \theta(z_0)\right)$$

$$E = \rho C_V U \left(q - q(z_0)\right)$$

- 1^{st} order parameterization
  - The fluxes are computed as function of mean quantities
- The Drag and exchange coefficients $C_{D,H,V}$ are function of atmospheric stability, quantified in terms of Richardson number:

$$Ri = \frac{g}{T_V} \frac{\frac{\partial \theta}{\partial z}}{\left|\frac{\partial V}{\partial z}\right|^2} \approx \frac{g}{T_V} \frac{(\theta_1 - \theta_0)(z_1 - z_0)}{(u_1 - u_0)^2 (v_1 - v_0)^2}$$

Transfer coefficients

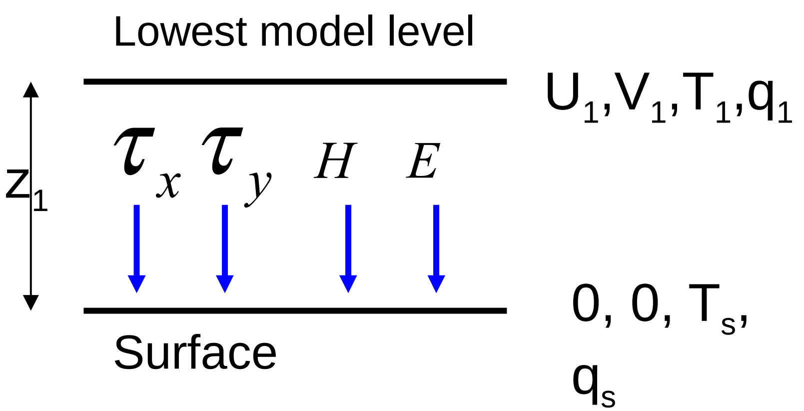
Surface fluxes can be written explicitly as:

$$\tau_x = \rho C_M |U_1| U_1$$

$$\tau_y = \rho C_M |U_1| V_1$$

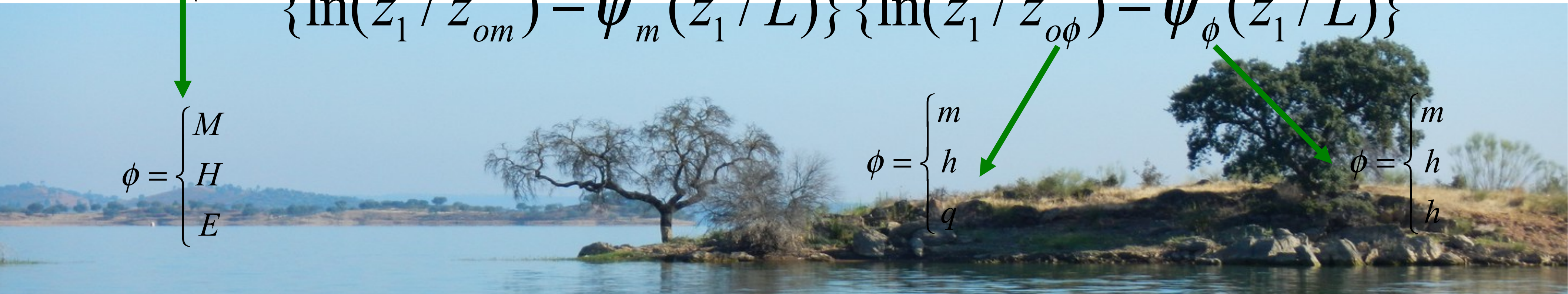
$$H = \rho c_p C_H |U_1| (\theta_1 - \theta_s)$$

$$E = \rho C_E |U_1| (q_1 - q_s)$$



where $|U_1| = \left(U_1^2 + V_1^2\right)^{1/2}$

and $$C_\phi = \frac{k^2}{\{\ln(z_1/z_{om}) - \psi_m(z_1/L)\}\{\ln(z_1/z_{o\phi}) - \psi_\phi(z_1/L)\}}$$



$$\phi = \begin{cases} M \\ H \\ E \end{cases} \qquad \phi = \begin{cases} m \\ h \\ q \end{cases} \qquad \phi = \begin{cases} m \\ h \\ h \end{cases}$$

Numerical procedure: The Richardson number

The expressions for surface fluxes are implicit i.e they contain the Obukhov length which depends on fluxes. The stability parameter z/L can be computed from the bulk Richardson number by solving the following relation:

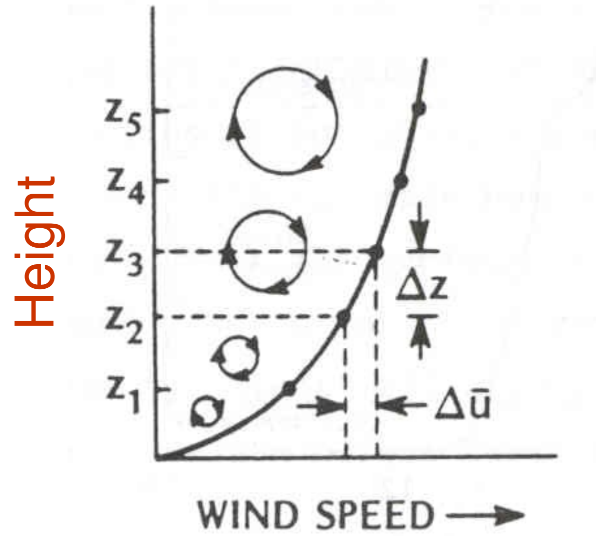
$$Ri_b = \frac{g z_1}{\theta} \frac{\theta_1 - \theta_s}{|U_1|^2} = \frac{z_1}{L} \frac{\{\ln(z_1 / z_{oh}) - \psi_h(z_1 / L)\}}{\{\ln(z_1 / z_{om}) - \psi_m(z_1 / L)\}^2}$$

This relation can be solved:

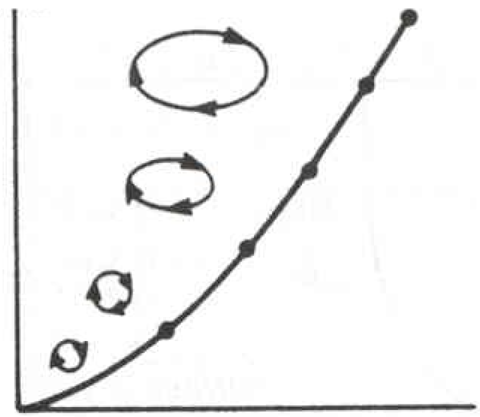
- Iteratively;
- Approximated with empirical functions;
- Tabulated.

U-Profile … Effects of Stability

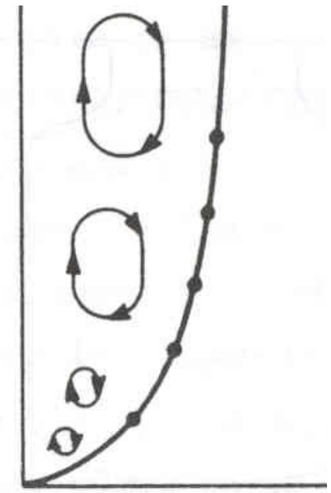
Neutral



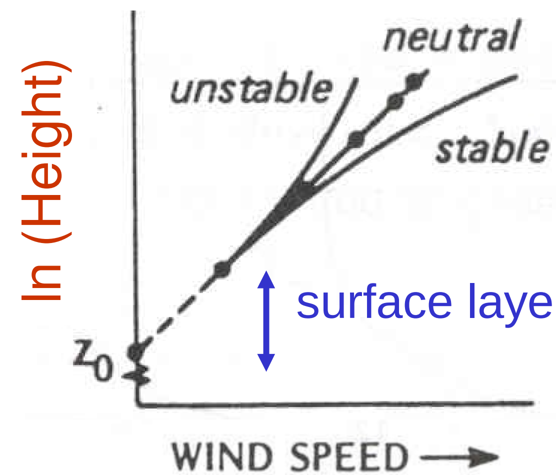
Stable



Unstable



Neutral: $$U = \frac{u^*}{\kappa} \ln \frac{z}{z_0}$$

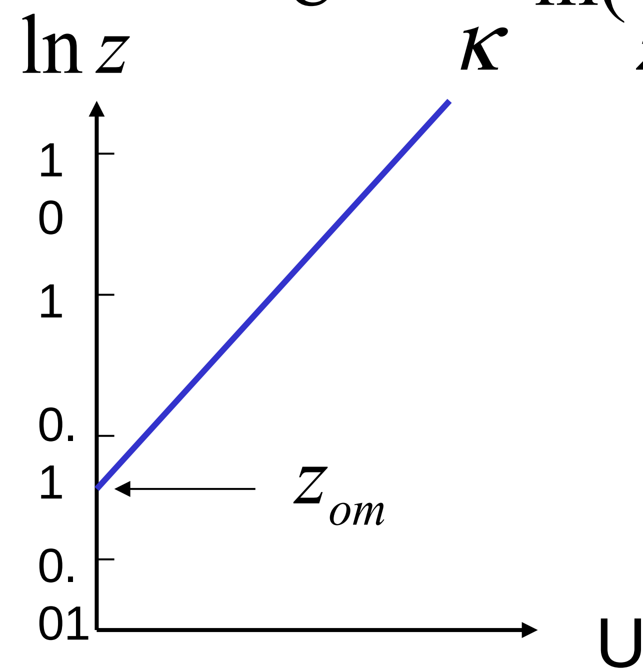


Oke 1978

Surface roughness length (definition)

Example for wind:

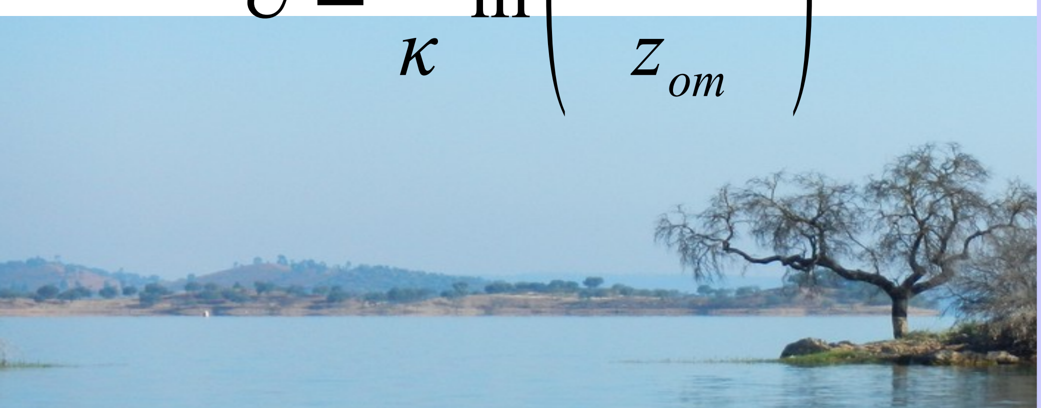
$$U = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_{om}}\right)$$



- Surface roughness length is defined on the basis of logarithmic profile.
- For z/L small, profiles are logarithmic.
- Roughness length is defined by intersection with ordinate.

Often displacement height is used to obtain U=0 for z=0:

$$U = \frac{u_*}{\kappa} \ln\left(\frac{z + z_{om}}{z_{om}}\right)$$



- Roughness lengths for momentum, heat and moisture are not the same.
- Roughness lengths are surface properties.

Roughness length over land

Ice surface	**0.0001 m**
Short grass	**0.01 m**
Long grass	**0.05 m**
Pasture	**0.20 m**
Suburban housing	**0.6 m**
Forest, cities	**1-5 m**

Roughness lengths over water

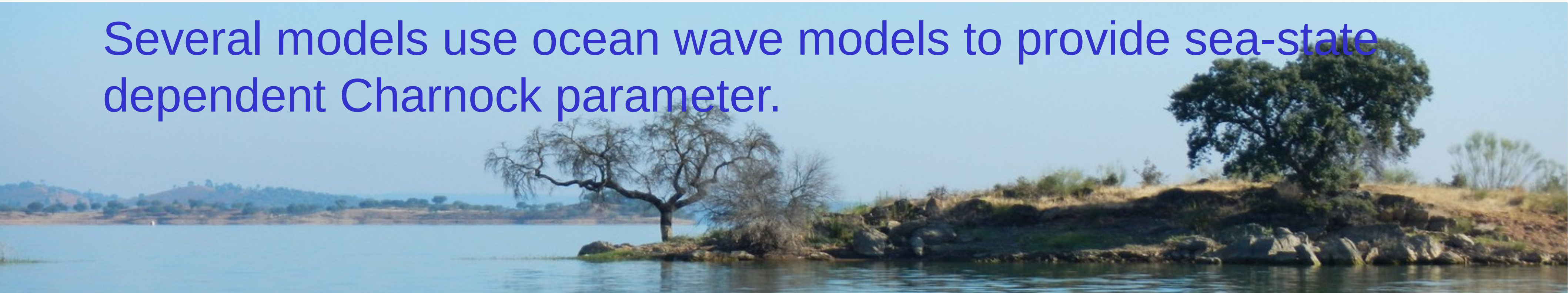
Roughness lengths are determined by molecular diffusion and ocean wave interaction e.g.

$$z_{om} = C_{ch}\frac{u_*^2}{g} + 0.11\frac{\nu}{u_*}, \; C_{ch} \;\; is \;\; Charnock \;\; parameter$$

$$z_{oh} = 0.40\frac{\nu}{u_*}$$

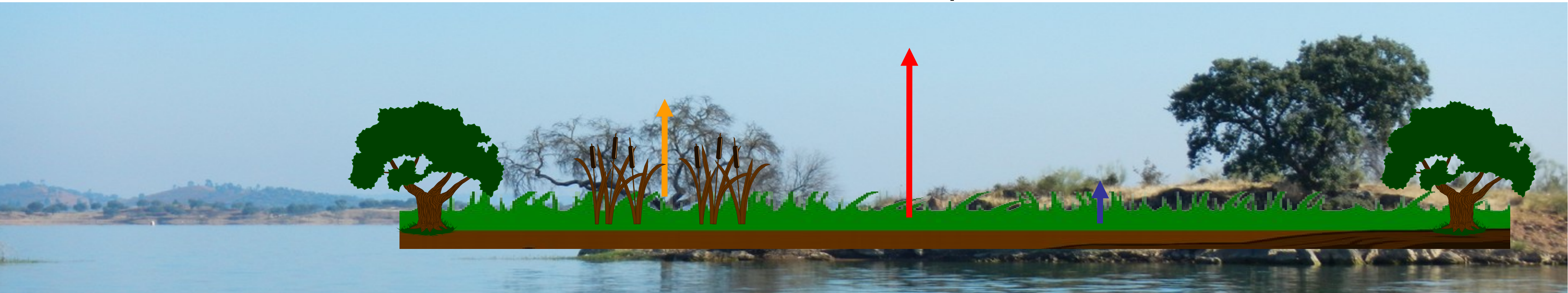
$$z_{oq} = 0.62\frac{\nu}{u_*}$$

Several models use ocean wave models to provide sea-state dependent Charnock parameter.



Effects of vegetation

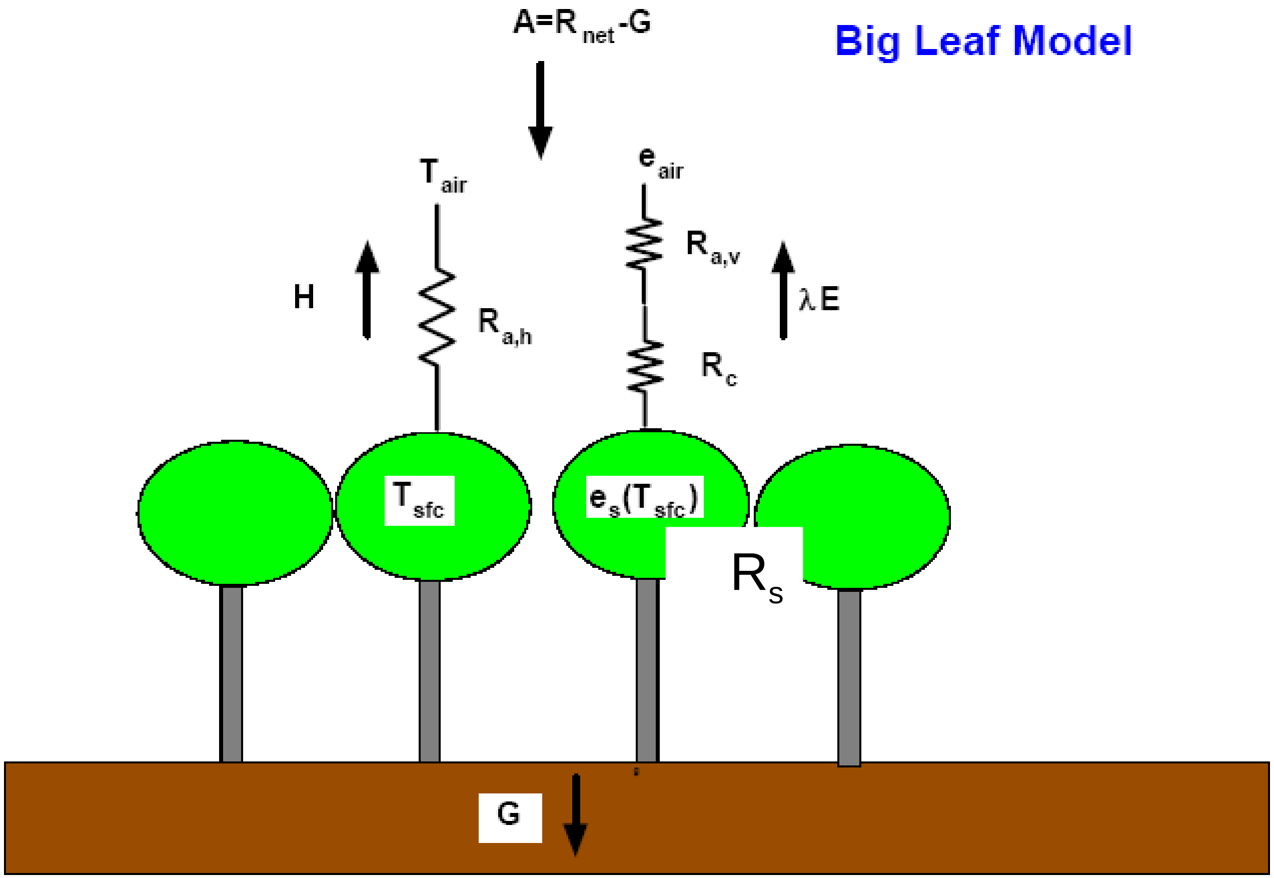
- To solve the equations of energy and water conservation we need to represent the vegetation.
- A realist formulation should include the effects of vegetation on:
  - evaporation, including the transpiration,
  - the surface heat flux partition,
  - the interception of precipitable water,
  - the soil water content,
  - the radiative proprieties of the surface,
  - the aerodinamic rugosity.
- From the meteorological point of view, the plants are the more efficient mechanism to transfer water from the soil to the atmosphere.
- The plants have physiological mechanisms that allow them to adapt to environmental conditions and control the transpiration rate.



Parameterization of the vegetation

- There is possible to study the energy fluxes inside the vegetation cover, but in meteorology and in NWP models the state of the art is to consider one layer models
  - big leaf aproximation
- They use the stomatal resistance as a critical quantity, which is aggregated to the considered scale. The complexity of the models depends on the processes and quantities considered to compute it.

Big Leaf Aproximation



Transpiration: The big leaf approximation

- Sensible heat (*H*), the resistance formulation

$$H = \rho C_p C_h u L \left(T_L - T_{sk}\right) = \rho C_p \frac{T_L - T_{sk}}{r_a}$$

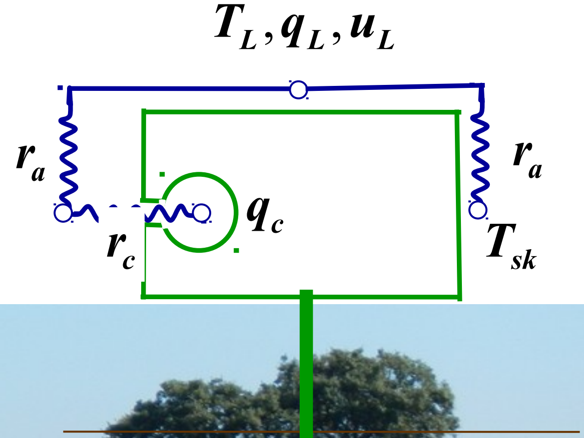
r_a ,aerodynamic resistance, $\left[r_a\right] = s\,m^{-1}$ $r_a = \dfrac{1}{C_h u_L}$

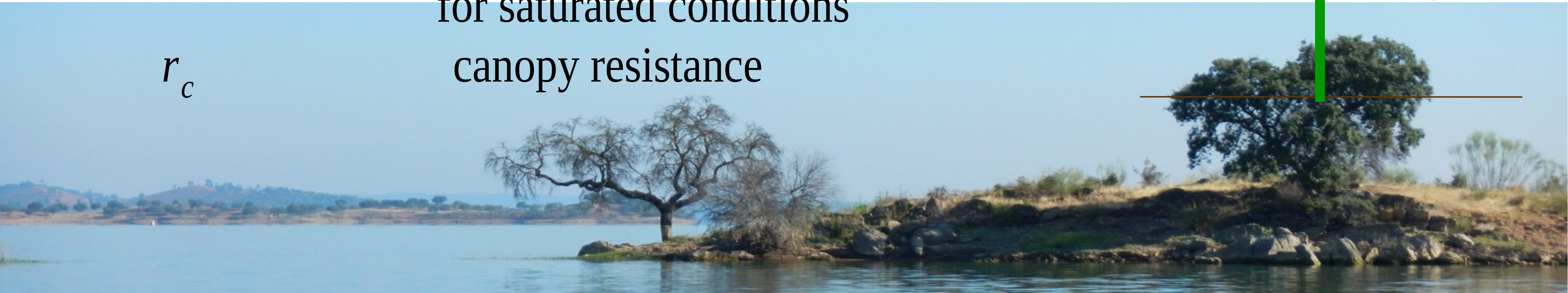
- Evaporation (*E*), the resistance formulation (the big leaf approximation, Deardorff 1978, Monteith 1965)

$$E = \rho \frac{q_L - q_c}{r_a + r_c} = \rho \frac{q_L - q_{sat}\left(T_{sk}\right)}{r_a + r_c}$$

$q_c = q_{sat}\left(T_{sk}\right)$ Specific humidity for the interior of the stomata, ie, for saturated conditions

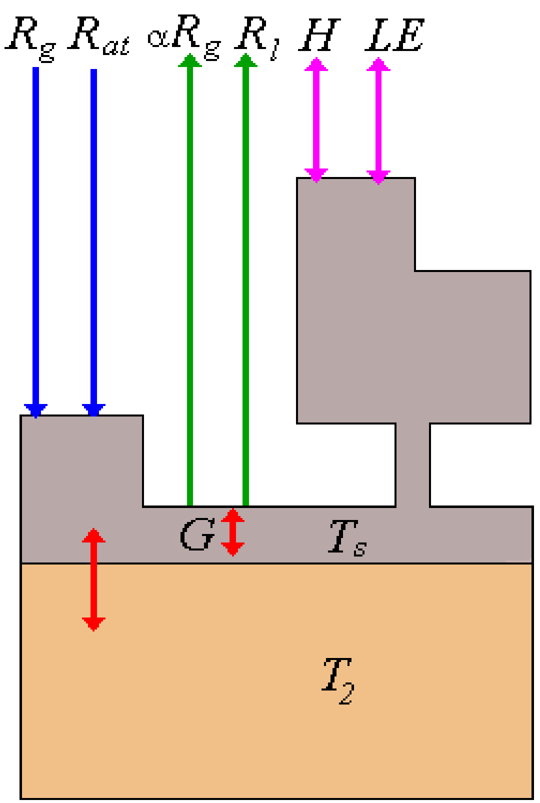
r_c canopy resistance





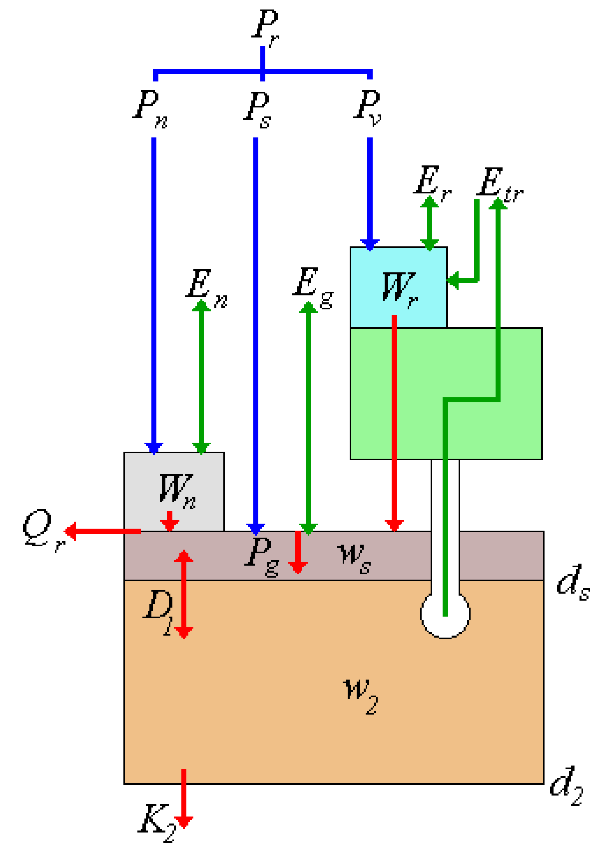
Vegetation parameterization in NWP models

Example: ISBA Model – Météo-France (included in the SURFEX platform, used by ALADIN & HIRLAM consortiums



Energy Budget

Only one Ts, for soil and cover.



Water Budget

TESSEL scheme

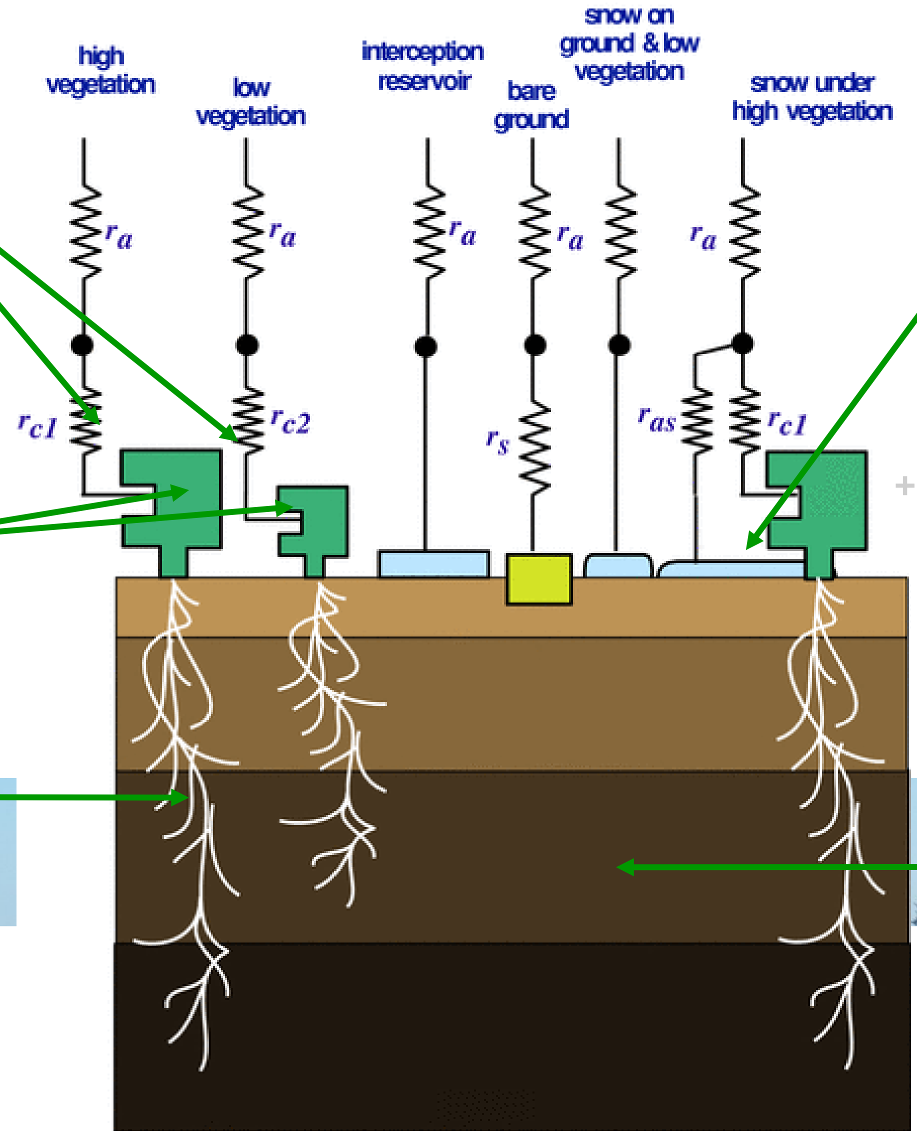
- Tiled ECMWF Scheme for Surface Exchanges over Land

Land surface tiles in ERA40 surface scheme

Revised canopy resistances, including air humidity stress on forest

High and low vegetation treated separately

Variable root depth



+ 2 tiles (ocean & sea-ice)

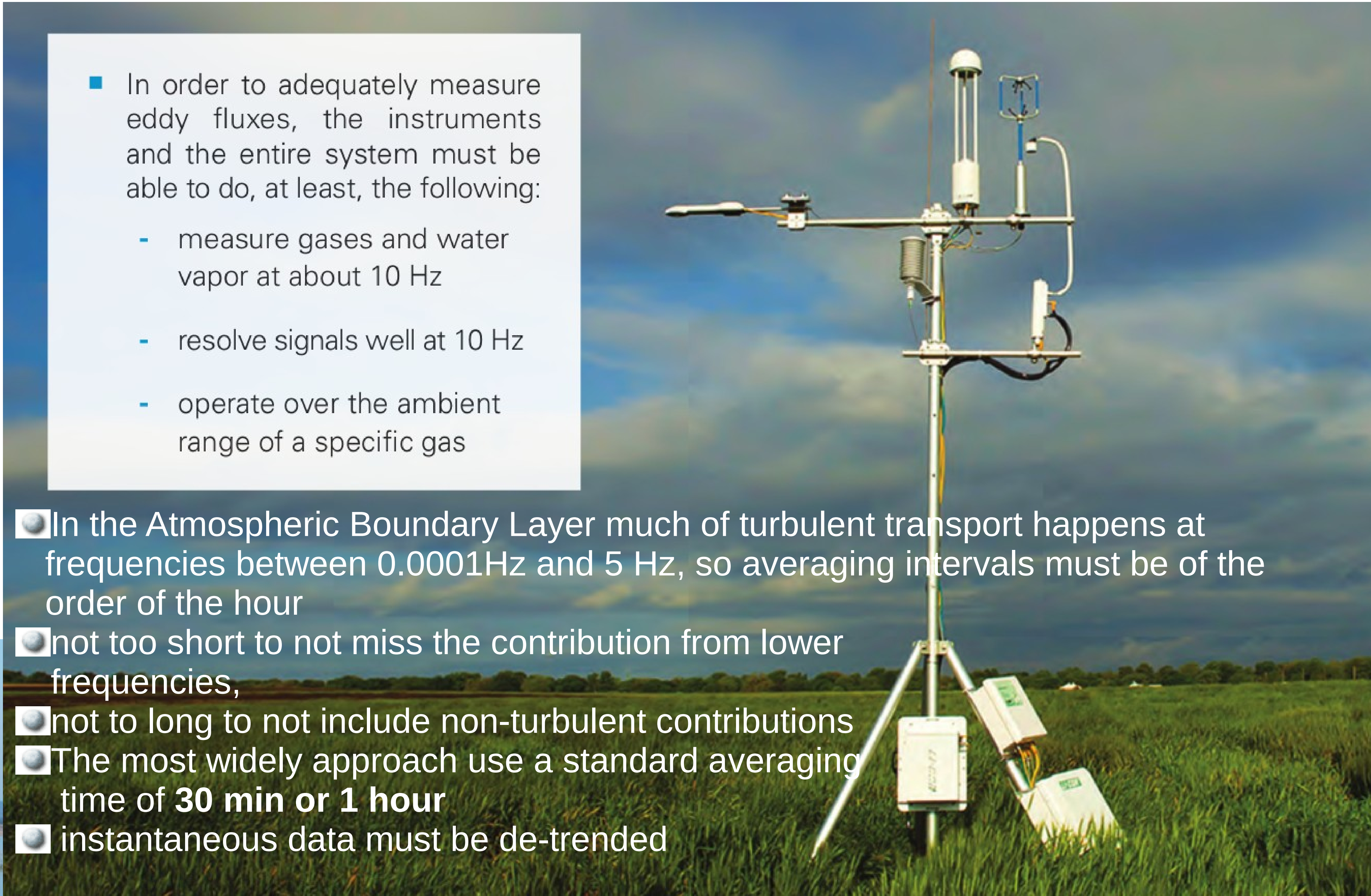
New treatment of snow under high vegetation

No root extraction or deep percolation in frozen soils

How to measure turbulent fluxes?

- The period of the eddies depend on its size: smaller eddies rotate faster – higher frequency
- There is always a mix of different eddy sizes, so turbulent transport is done at different frequencies: from large movements of the order of hours to small ones on the order of 1/10 second.
- So, instruments need to be fast
- Covariances must be computed over a relatively long period

How to measure and compute Fluxes



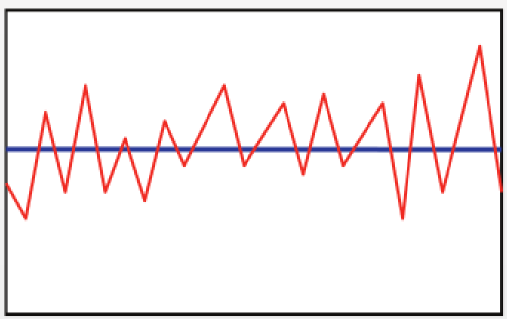
- In order to adequately measure eddy fluxes, the instruments and the entire system must be able to do, at least, the following:
  - measure gases and water vapor at about 10 Hz
  - resolve signals well at 10 Hz
  - operate over the ambient range of a specific gas

- In the Atmospheric Boundary Layer much of turbulent transport happens at frequencies between 0.0001Hz and 5 Hz, so averaging intervals must be of the order of the hour
- not too short to not miss the contribution from lower frequencies,
- not to long to not include non-turbulent contributions
- The most widely approach use a standard averaging time of **30 min or 1 hour**
- instantaneous data must be de-trended

Data de-trending

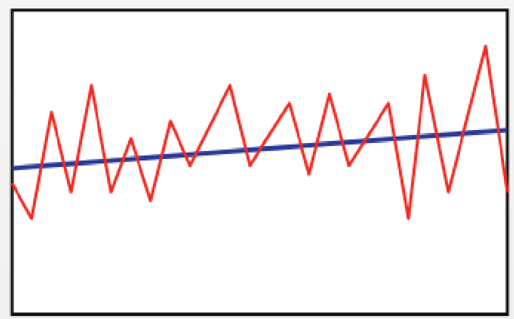
- Mean values are subtracted from instantaneous values to compute flux
- This requires establishing the mean for a given time series
- There are three main ways to look at it, and three respective techniques

Block averaging
(mean removal)



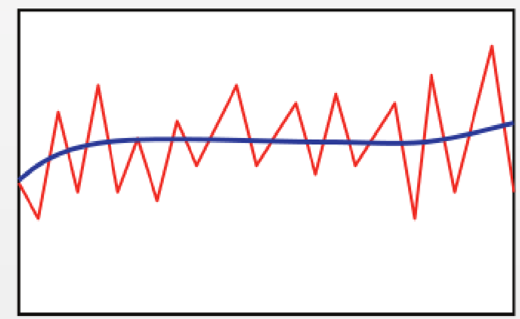
- Simplest situation
- Many prefer this method
- May gain artificial flux

Linear detrending
(linear trend removal)

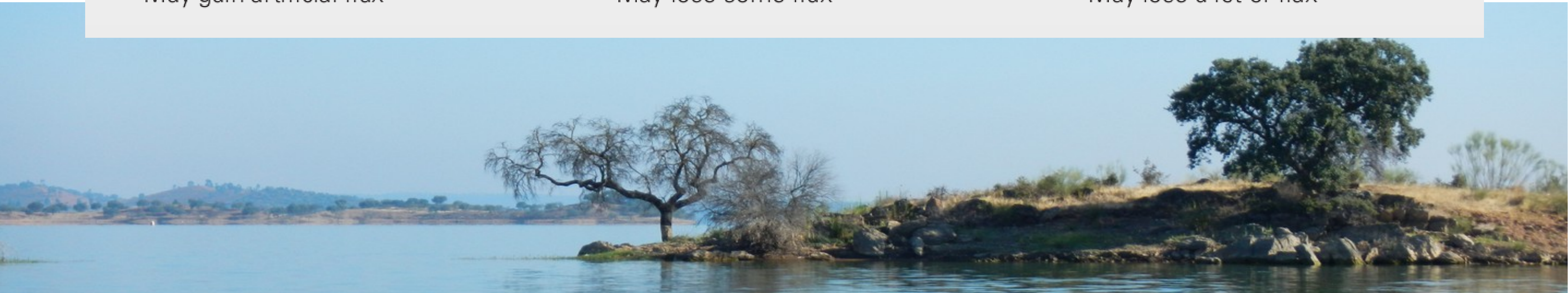


- For example, sensor drifts
- Rapid diurnal changes
- May lose some flux

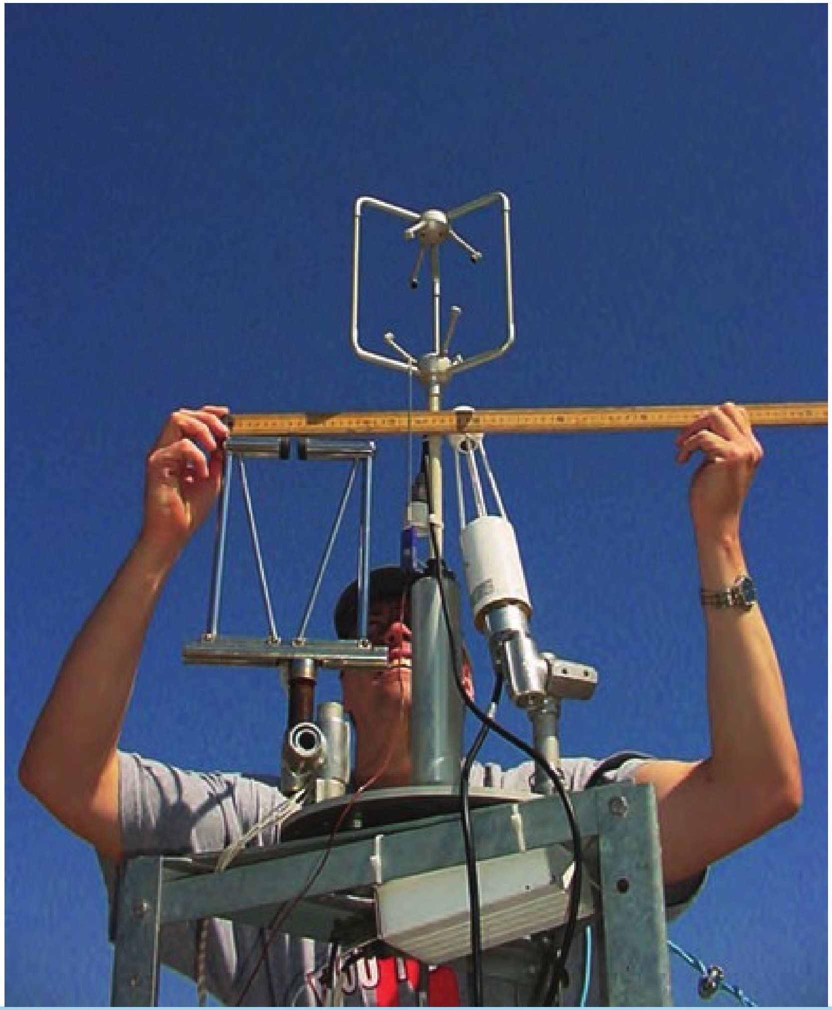
Non-linear filtering
(non-linear trend removal)



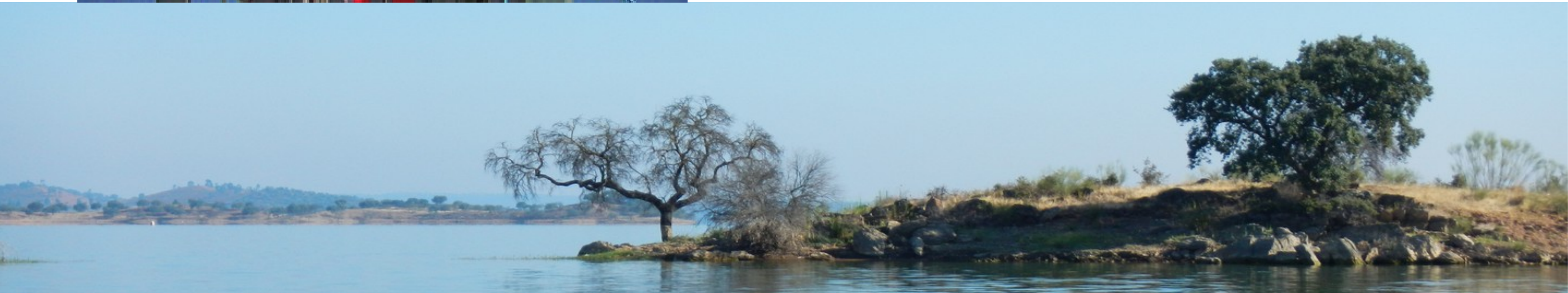
- Complex situation
- Same as high pass filter
- May lose a lot of flux



How to compute Fluxes



- sonic anemometers are the key elements
- sonic anemometer cannot be perfectly leveled – a coordinate rotation must be applied to ensure that w=0
- Other two rotations may be applied:
  - Align x direction to the mean direction (v=0)
  - impose $\overline{v'w'}=0$



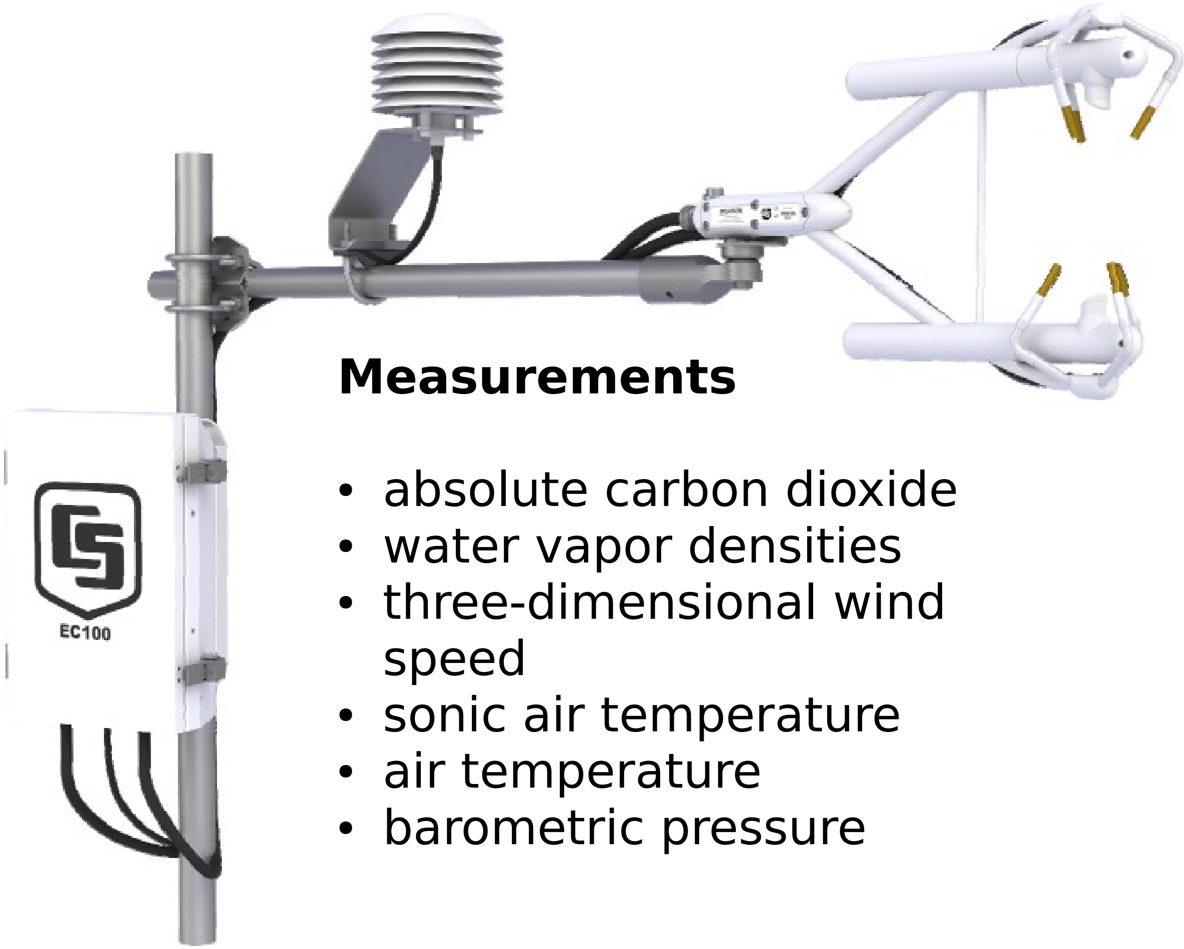
How to compute Fluxes: Mayor assumptions

- Measurements at a point can represent an upwind area
- Measurements are done inside the boundary layer of interest
- Fetch/footprint is adequate – fluxes are measured from the area of interest
- Flux is fully turbulent – most of the net vertical transfer is done by eddies
- Terrain is horizontal and uniform: average of fluctuations of w' is zero, air density fluctuations, flow convergence and divergence are negligible
- Instruments can detect very small changes at high frequency
- Air flow is not distorted by the installation structure or the instruments

Placement of the captures (examples)

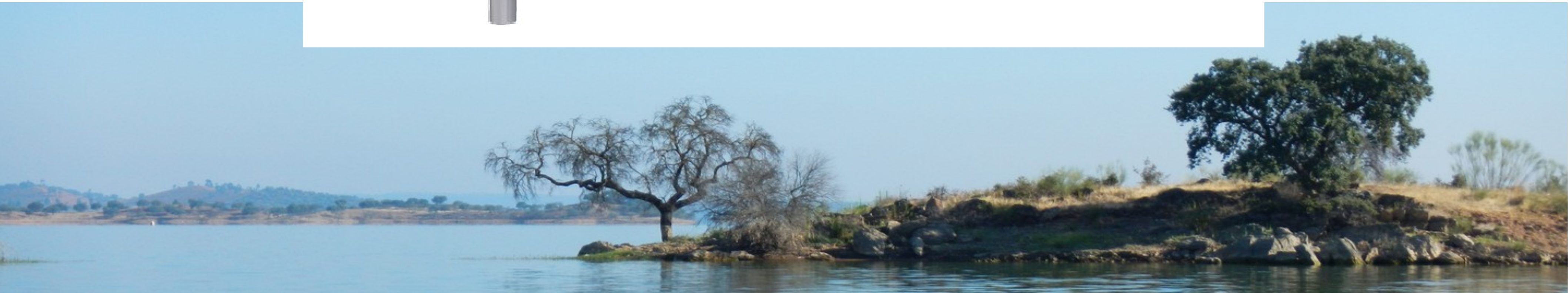
irgason

Integrated CO_2/H_2O Open-Path Gas Analyzer and 3D Sonic Anemometer

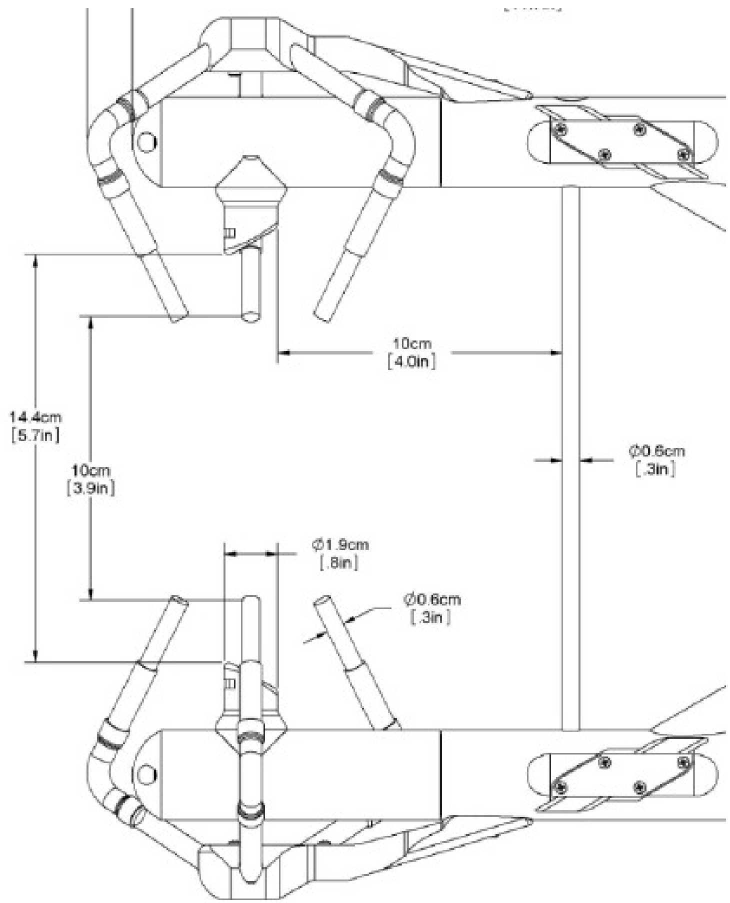


Measurements

- absolute carbon dioxide
- water vapor densities
- three-dimensional wind speed
- sonic air temperature
- air temperature
- barometric pressure



Measurement of Wind Speed

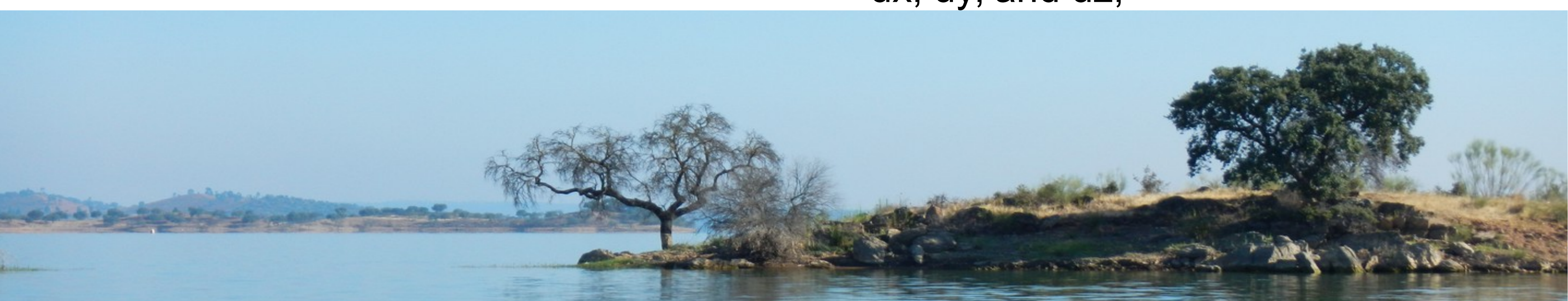


- Each axis of the sonic anemometer pulses two ultrasonic signals in opposite directions.
- The times of flight of the first signal (t_0) and of the second signal (t_b) allows the determination of the wind speed along the axis, u_a

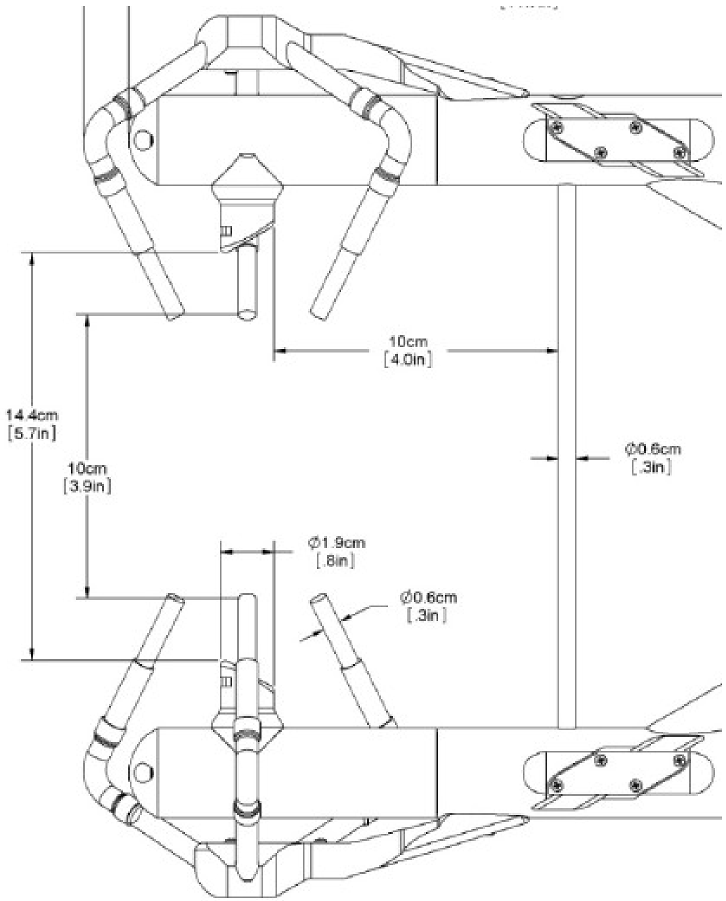
$$u_a = \frac{d}{2}\left[\frac{1}{t_o} - \frac{1}{t_b}\right]$$

(d is the distance between the transducers)

- The non-orthogonal wind speed components are then transformed into orthogonal wind speed components, ux, uy, and uz,



sonic virtual temperature



- The sonically determined speed of sound can be found from:

$$c = \frac{d}{2}\left[\frac{1}{t_o} + \frac{1}{t_b}\right]$$

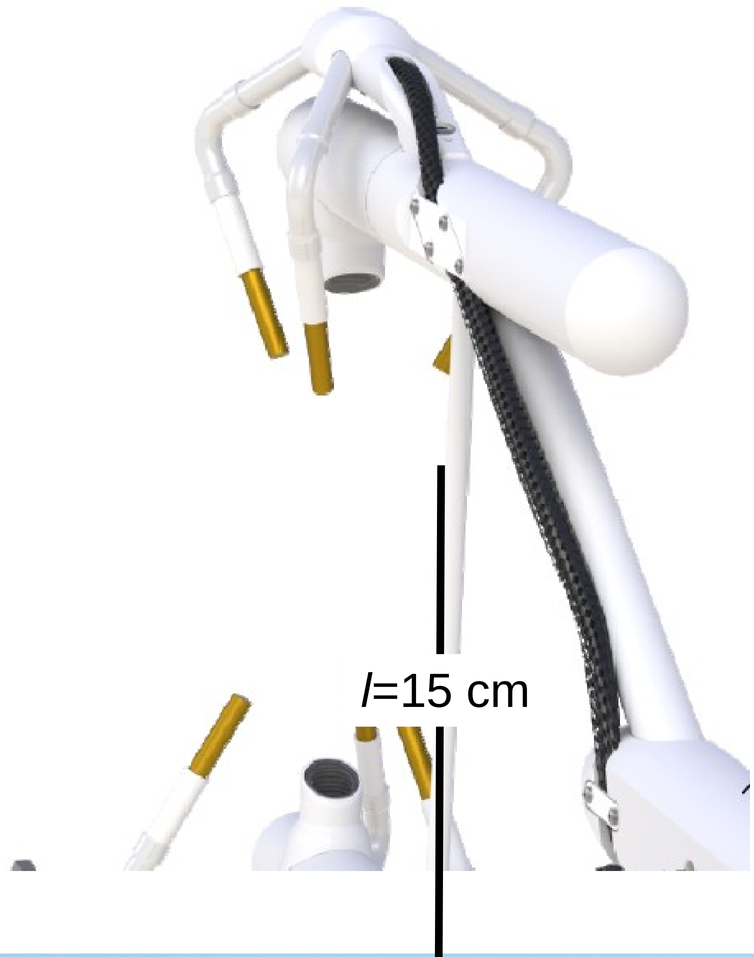
- The speed of sound in moist air is a function of temperature and humidity and is approximatively given by:

$$c^2 = \gamma_d R_d T_s = \gamma_d R_d T(1 + 0.51q)$$

- where T_s is sonic virtual temperature, γ_d is the ratio of specific heat of dry air at constant pressure to that at constant volume and R_d is the gas constant for dry air

- Note: the effect of humidity, on the speed of sound, is included in the sonic virtual temperature

Gas analyzer

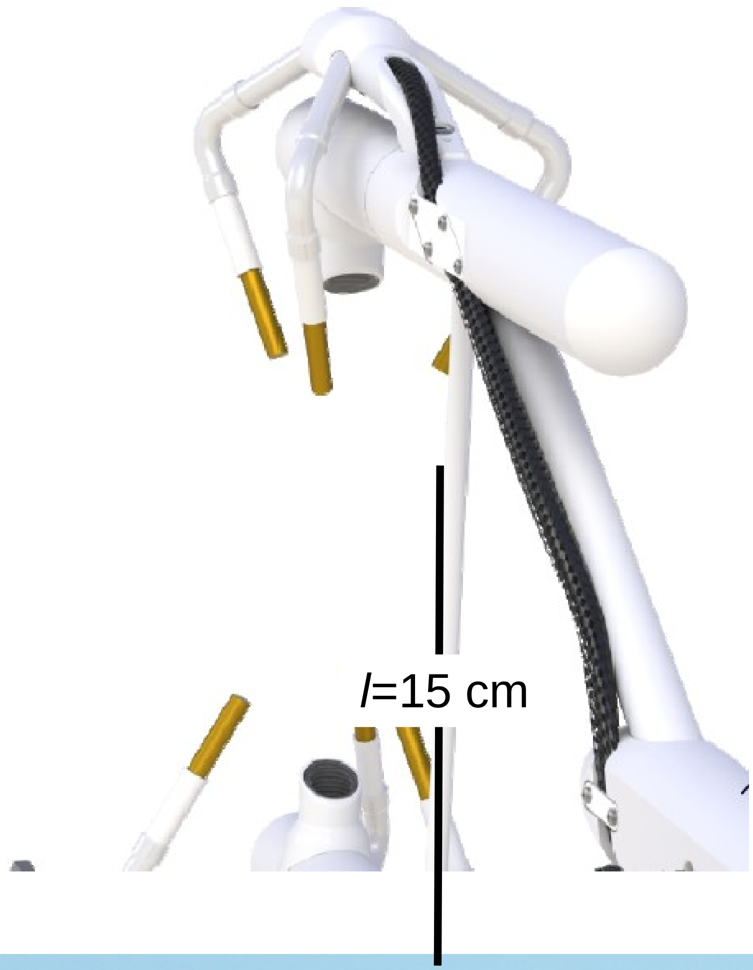


- The gas analyzer is a non-dispersive mid-infrared absorption analyzer.
- Infrared radiation is generated in the upper arm of the analyzer head before propagating along a 15 cm optical path.
- Chemical species located within the optical beam will absorb radiation.
- A mercury cadmium telluride detector in the lower arm measures the decrease in radiation intensity due to absorption, which can then be related to analyte concentration using the Beer-Lambert Law:

$$P = P_o e^{-\varepsilon c l}$$

P is irradiance after passing through the optical path, Po is initial irradiance, ε is molar absorptivity, c is analyte concentration, and l is pathlength.

Gas analyzer

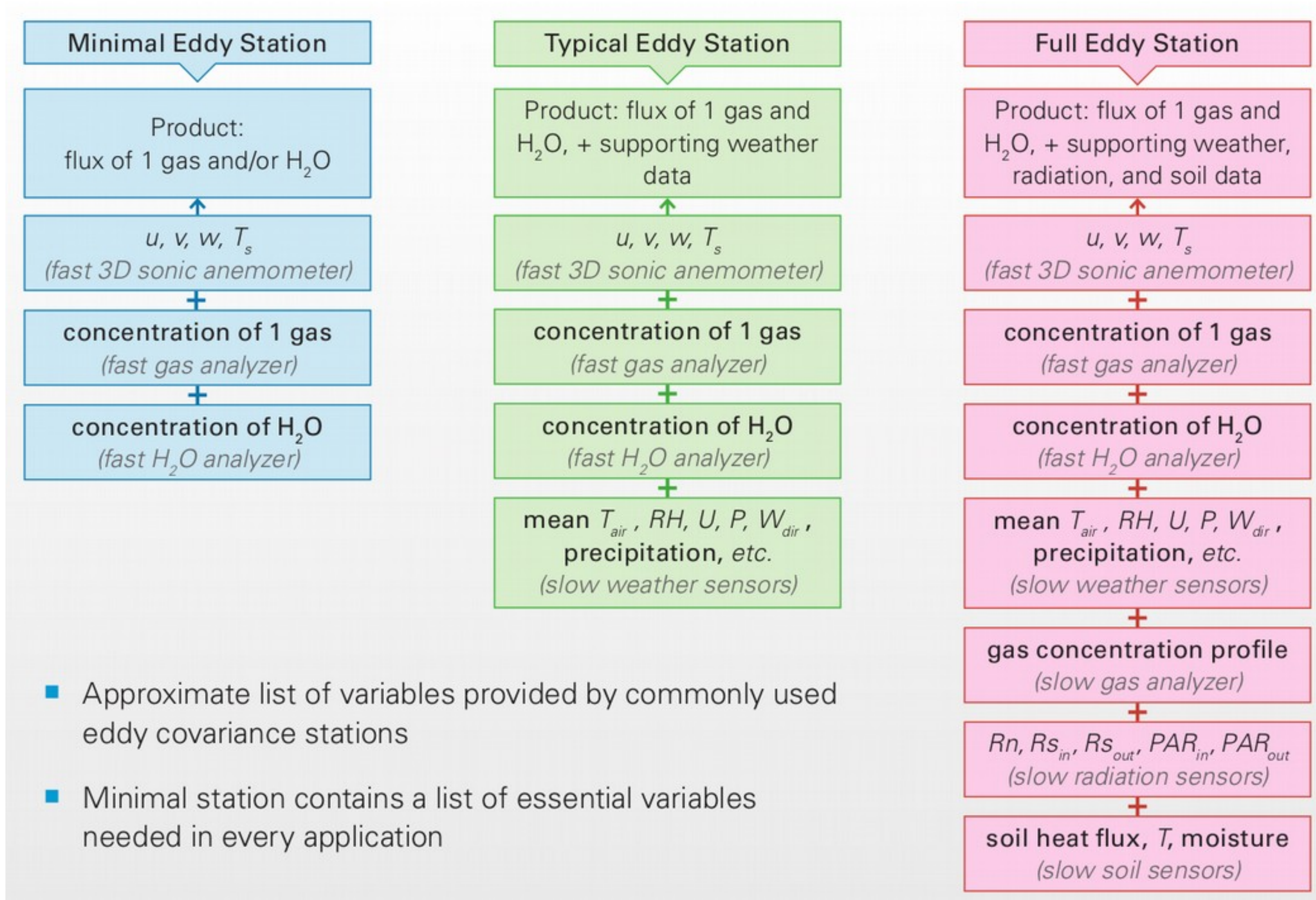


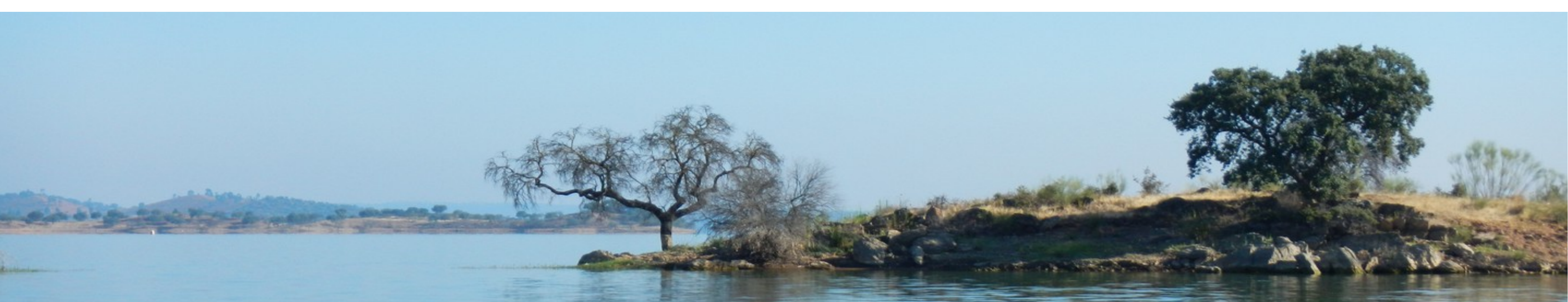
- For CO_2, light with a wavelength of 4.3 µm is selected - the molecule's asymmetric stretching vibrational band.
- For H_2O, radiation at 2.7 µm is used - water's symmetric stretching vibrational band

EC100

- The EC100 electronics digitize and process the detector data to give the CO2 and H2O densities after each chopper wheel revolution (100 Hz).
- The EC100 also synchronously measures and processes data from the sonic anemometer.

Eddy Station





Applications

- Emission rates from industries, landfills, *etc*.
- Carbon capture and sequestration
- Cap-and-trade compliance and quantification
- Line efficiencies and leak detection

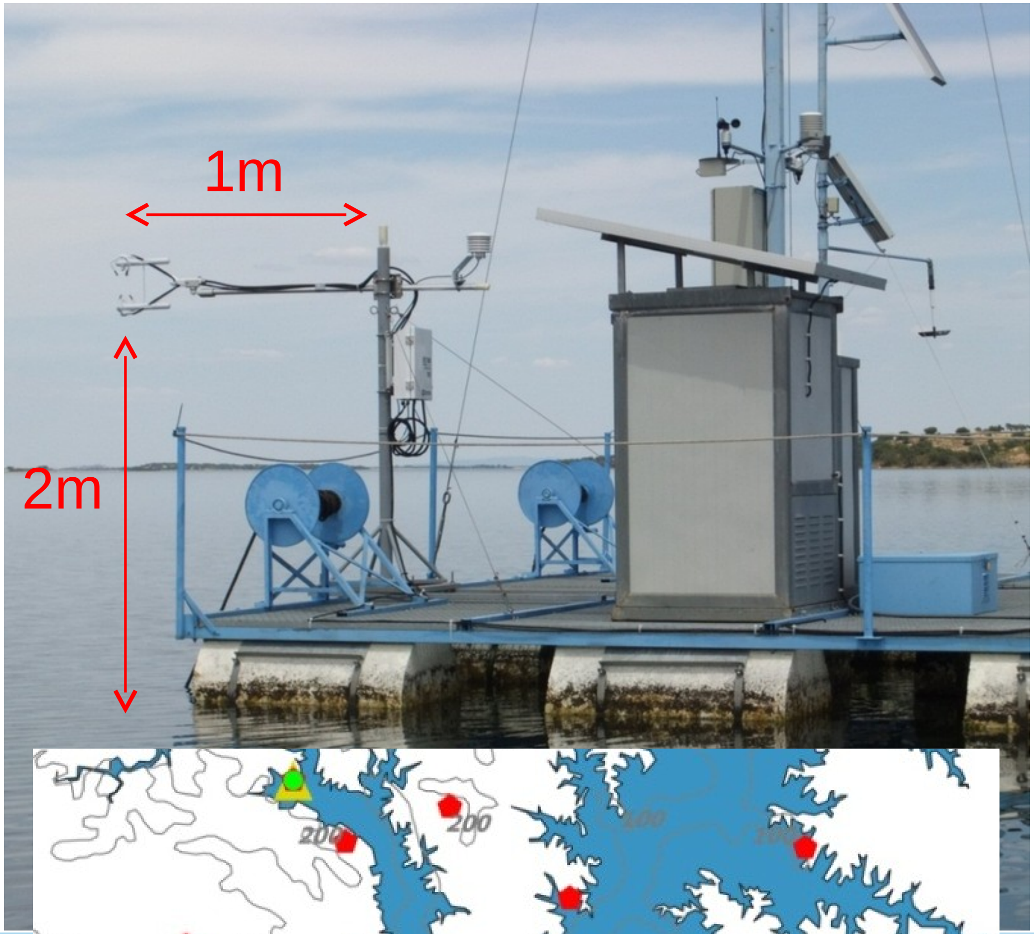
Applications

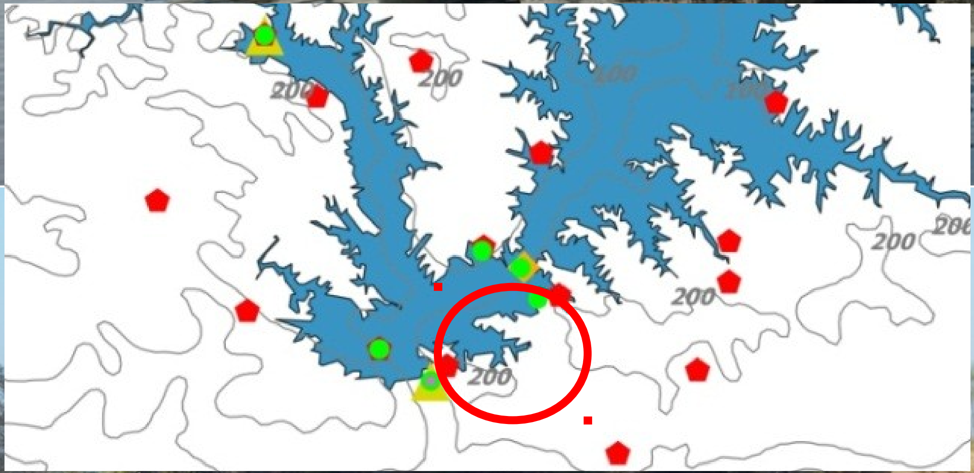
- High-precision irrigation
- Irrigation efficiency
- Water use and water use efficiency

Eddy covariance in Alqueva reservoir

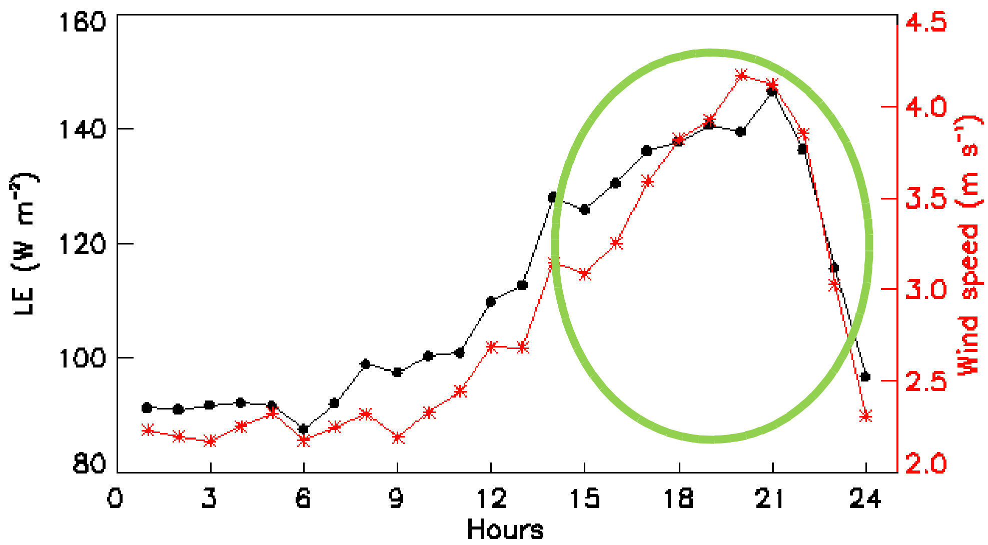
System: IRGASON
Frequency: 20 Hz
height: 2 m
Flux averages: 30 min
Orientation: Northwest (prevailing winds in summer caused by Iberian thermal low)

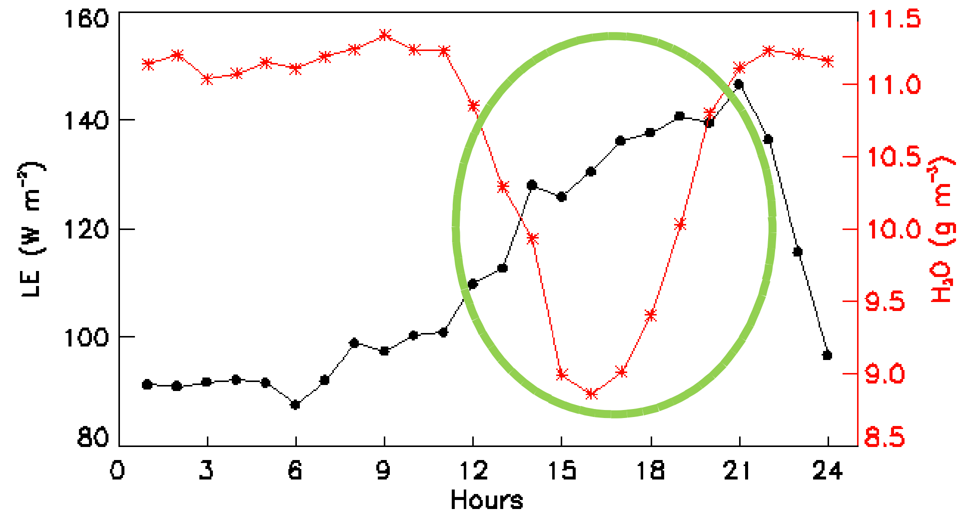
Built-in accelerometer in Waspmote board – Libelium to compute the vertical velocity of the arm



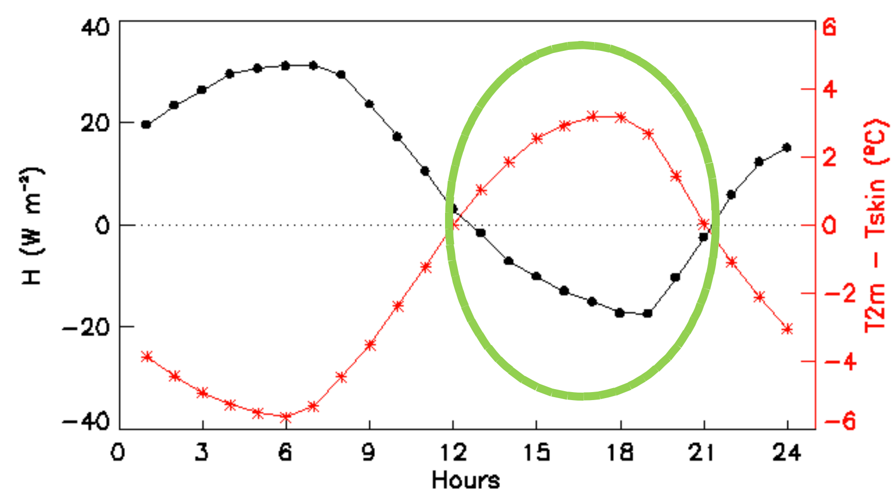


Heat fluxes (June to September)

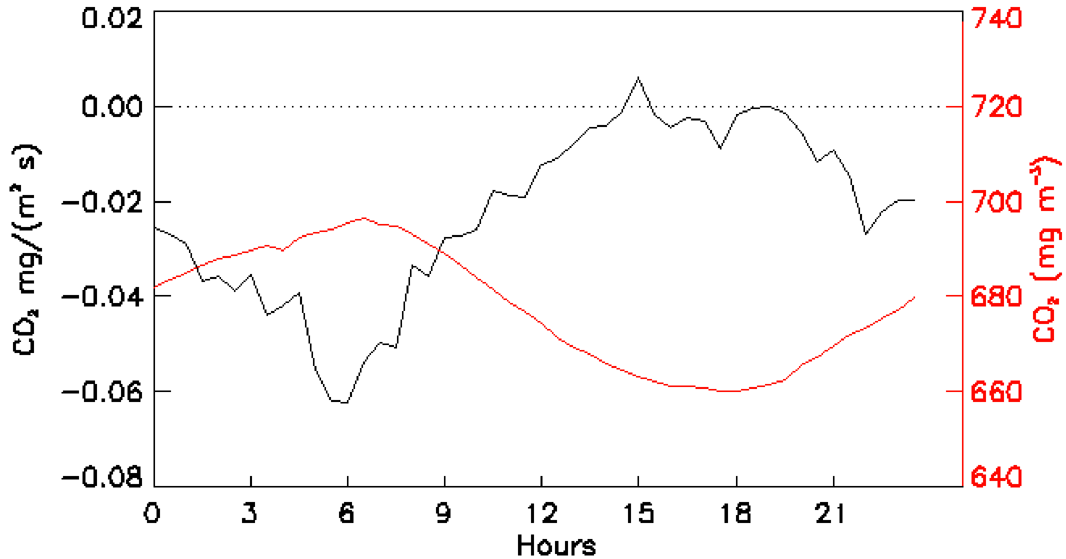




During the afternoon, between 12 and 21 hours, the air temperature is hotter then reservoir surface and lake breeze is developed allowing the subsidence of upper dry air leading to an increase of latent heat and forcing a negative sensible heat flux.



CO_2 Flux

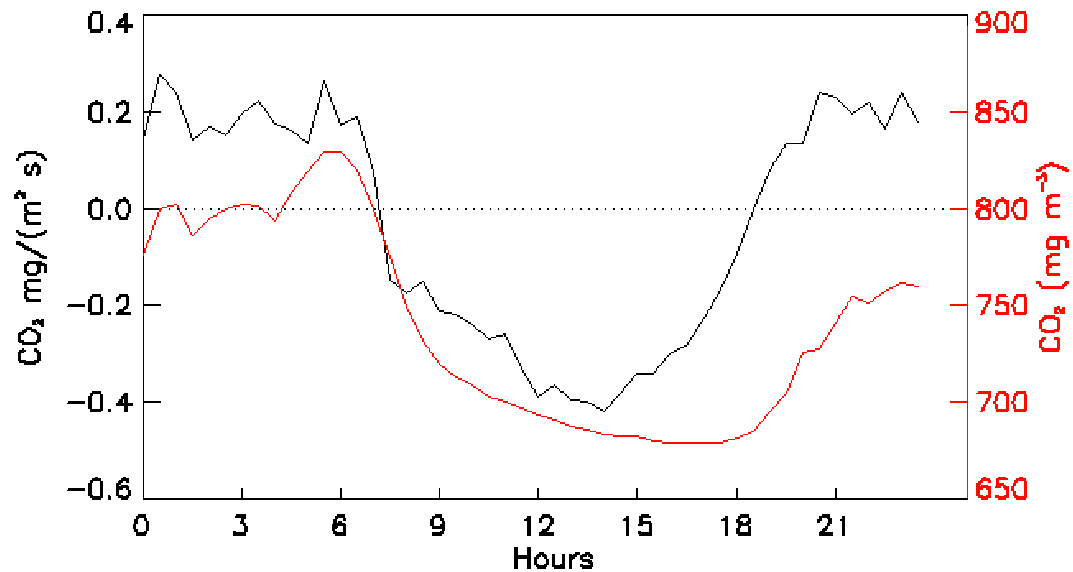


Over the reservoir (July)

When the CO_2 concentration is higher (night and morning) the flux is much negative than for lower concentration (during afternoon) thus dam is absorbing more CO_2 in this period.

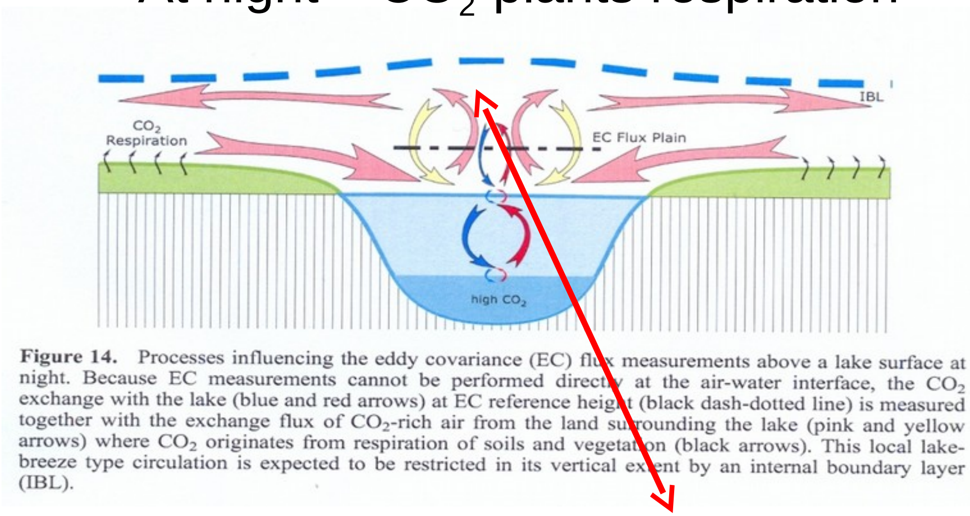
Over grass (April)

During April the system was mounted inland to perform the tests. The local was covered with grass. The results show positive flux during night and negative during daytime. The flux is correlated with the concentration.

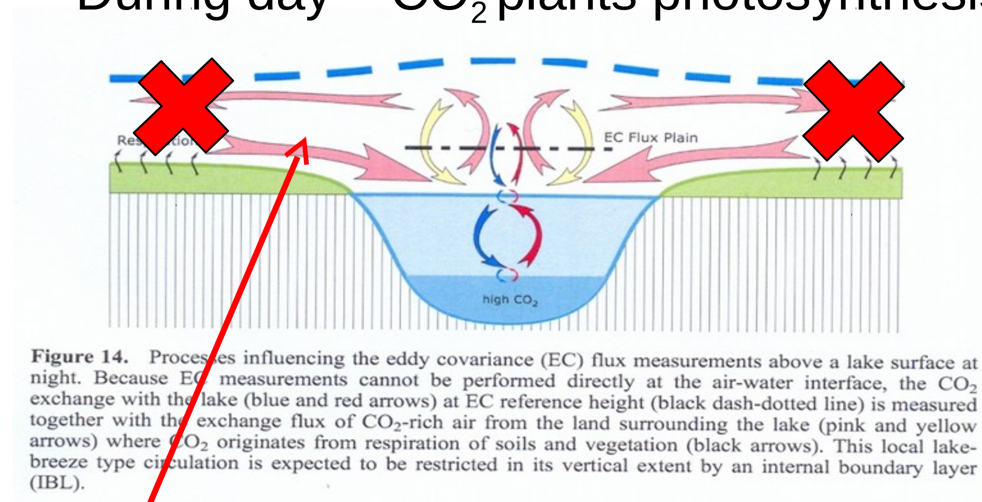


CO_2 over reservoir (June to September)

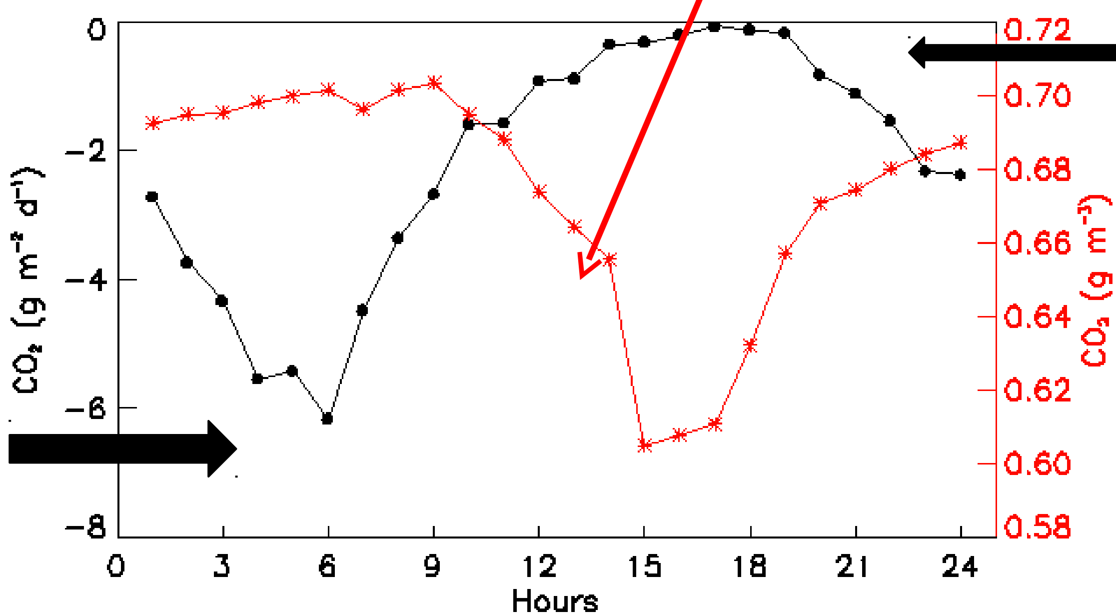
At night – CO_2 plants respiration



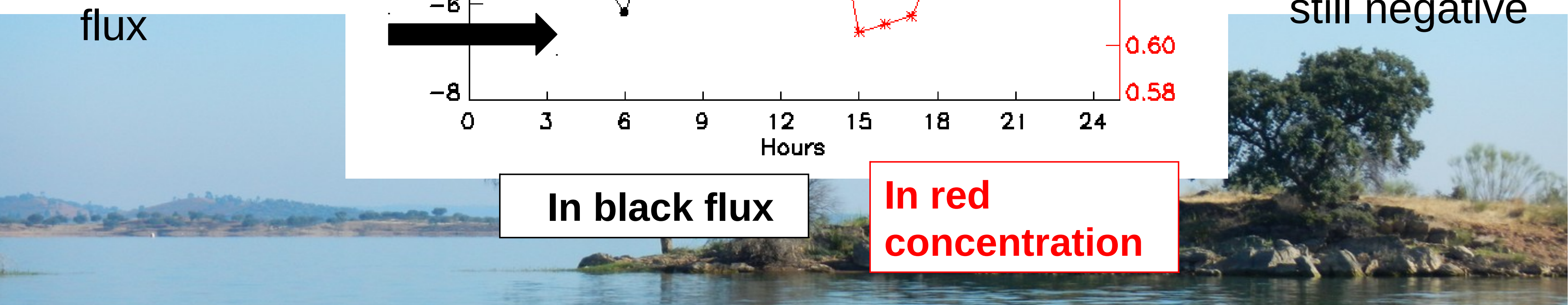
During day – CO_2 plants photosynthesis



Greater uptake by the reservoir during night – high negative flux



Lower uptake by the reservoir during day – weaker flux, still negative



In black flux

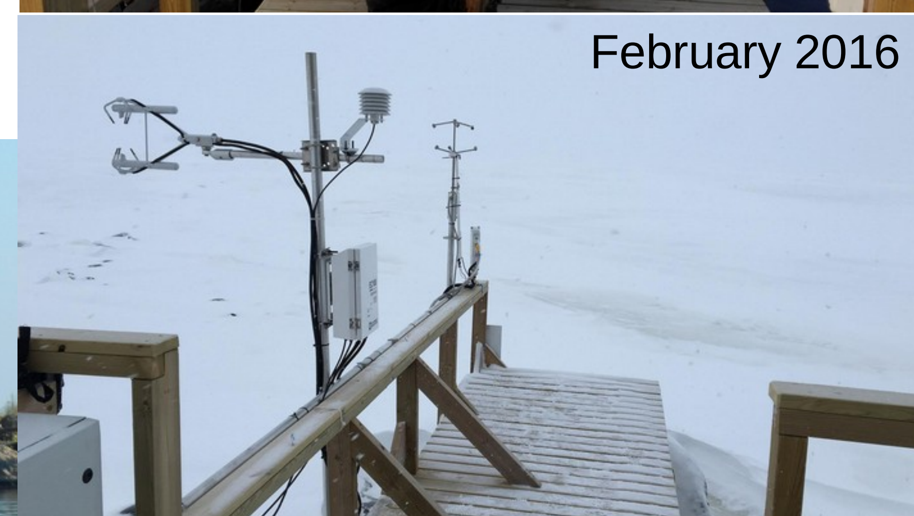
In red concentration

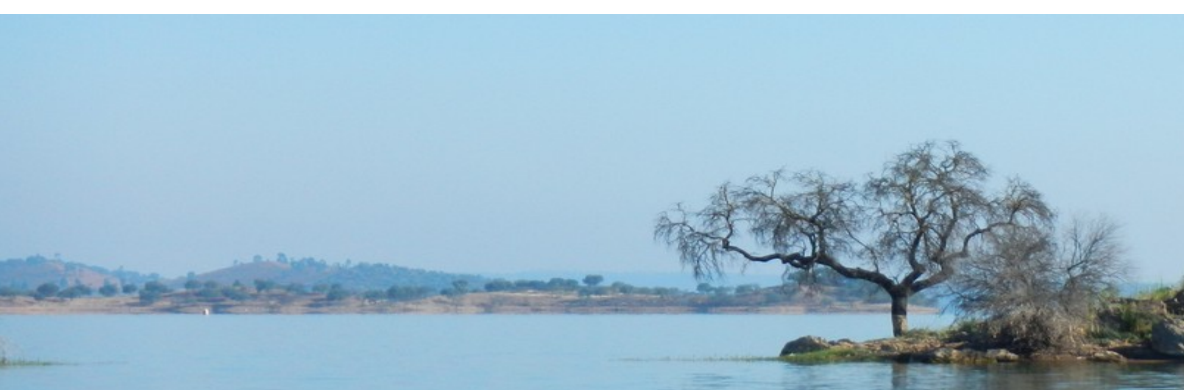
Eddy covariance Measurements over lake and snow in Finland

U. Évora / U. Helsinki inter comparison Experiment
Measurement site is located in a tip of narrow peninsula on the lake Vanajavesi, offering very good conditions for eddy covariance flux measurements.


Installation: 3 November 2015

- The EC system was installed at 2.5m height above the lake surface and was oriented against the prevailing wind direction in the site.
- The eddy-covariance system was installed in November 3, 2015 and is collecting data, continuously since then.
- Lake Vanajavesi started freezing over on 30 Dec, and it was completely frozen by 5 Jan

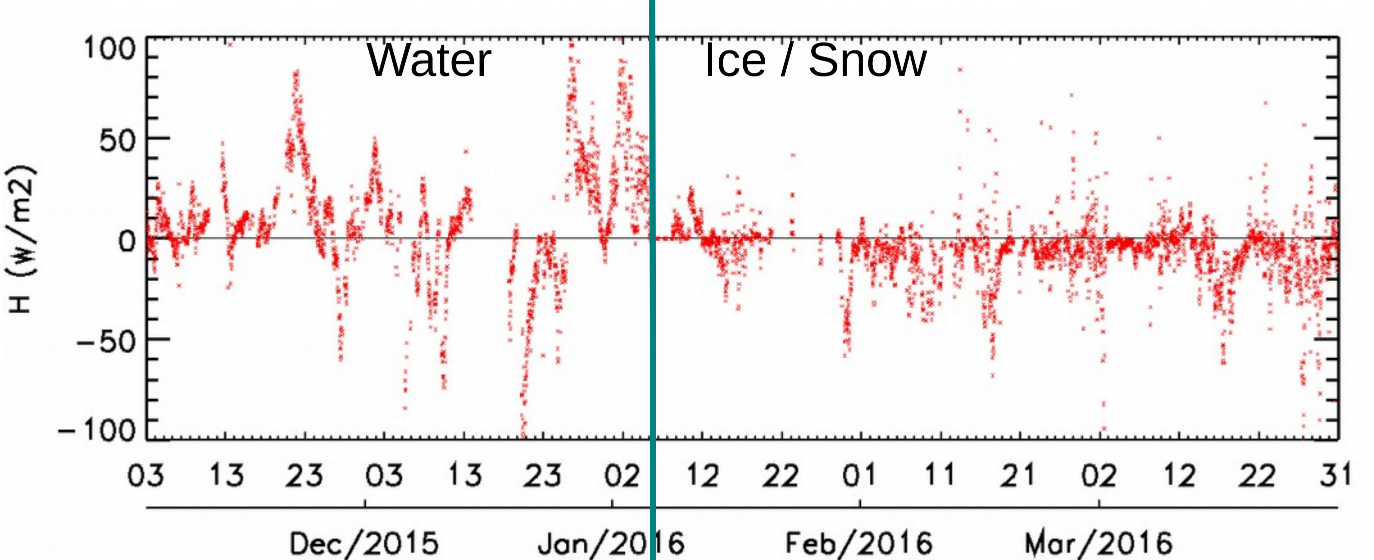

February 2016



Eddy covariance Measurements over lake and snow in Finland (results)

- As expected, sensible Heat fluxes (H) are very weak
- During freezing period H reaches 100 Wm-2
- Over ice/snow, H are in general between – 50 and 50 Wm-2
- In November, H is positive (water is losing energy), in March H is negative (air is heating the surface)
- Latent heat fluxes over ice/snow are very weak
- Nevertheless, in March the fluxes are slightly positive

Sensible Heat Flux



Latent Heat Flux