ETH
Eidgenössische Technische Hochschule Zürich
Swiss Federal Institute of Technology Zurich

In-situ measurements
Climatological and hydrological fieldwork FS2023

In this script a variety of meteorological instruments that are deployed at the hydro-meteorological research site Büel are shown. You will learn about their working principles and related terms (such as time constant, accuracy, precision).

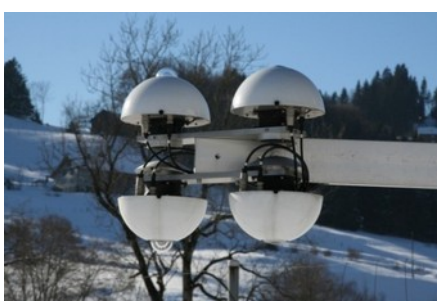
An overview on the current instrumentation can be found at: https://iac.ethz.ch/group/land-climate-dynamics/research/rietholzbach/instrumentation.html

1 Radiation

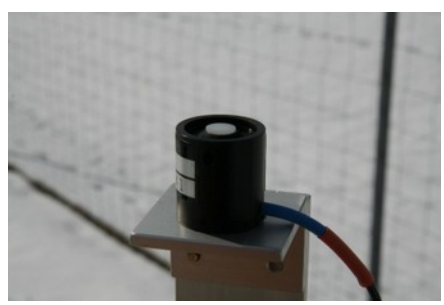
You will hear more about the radiation measurements in the corresponding experiment (e.g., what are the domes for? What is the ventilation for? What is the common measurement principle?). Here, an overview of the current instrumentation at Büel is given including the measured variable, the wavelength band, the manufacturer, and the sensor type:

Short-wave radiation 305-2800 nm
(Kipp&Zonen CM21)

Long-wave radiation 4.5-42 µm
(Kipp&Zonen CG4)

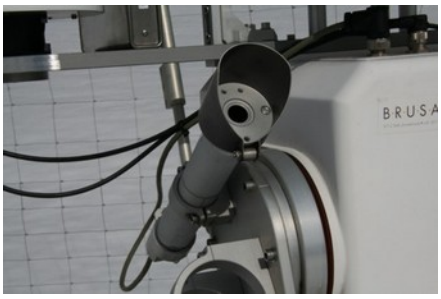


All four components of the radiation balance
(2x CM21, 2x CG4)



Photosynthetic active radiation (PAR)
400-700 nm (LiCor Li190SA)

Long-wave radiation 3.5-50 μm
(Eppley PIR)

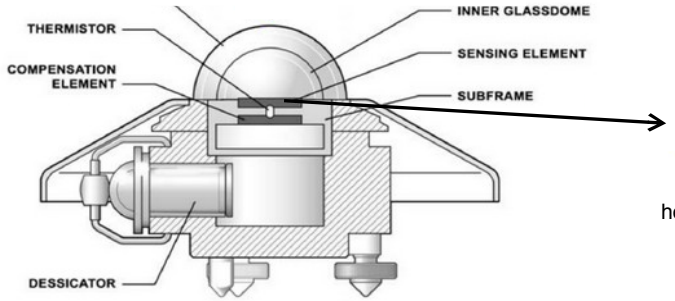


Direct short-wave radiation 0.4-4 μm
(Kipp&Zonen CH1)

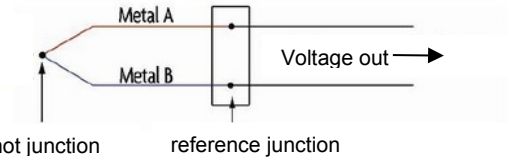
Sun tracker
(CM21, CH1, PIR)

Sunshine duration (threshold 120 W m^{-2})
(Meteoservis SD4)

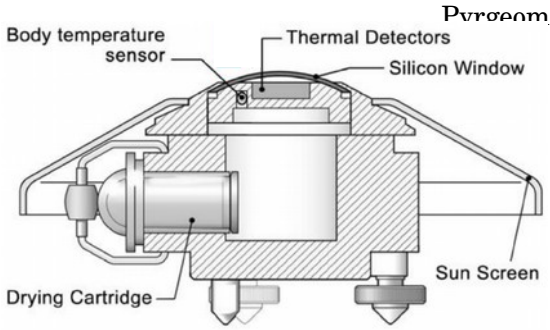
Pyranometer scheme:



Thermocouple



Pyrgeometer scheme:



2 Precipitation

Some questions:

- What are the main measurement principles?
- Are these measurements accurate or do they under-/overestimate the precipitation?
- What are sources of error with these instruments?

The instrumentation at Büel:

Tipping bucket @1.5m
(Lambrecht 1518-H3)

Electronic balance @1.5m
(OTT Pluviometer)

Tipping bucket @0m
(Gertsch Tognini)

Lysimeter @0m

3 Air Pressure

You will hear more about air pressure during the balloon sounding experiment. To capture the surface air pressure the instrument below is installed in the lysimeter cellar (see video):

Pressure sensor
(Vaisala PTB101B)

4 Wind Direction and Wind Speed

(Young 05103)

(Vaisala WMT52)

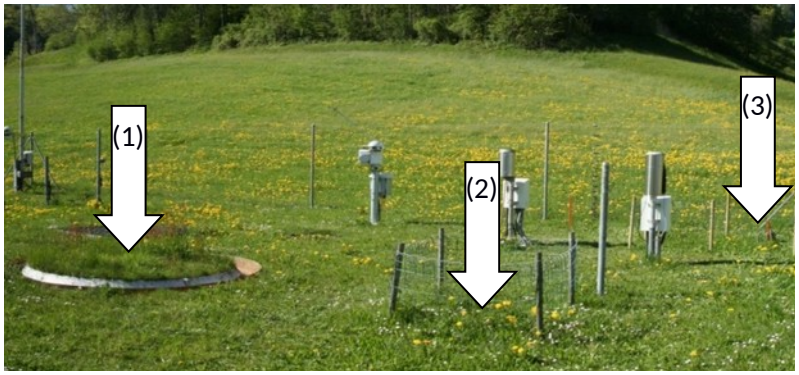
(Campbell CSAT3)

Wind direction and wind speed can be captured with different methods.

- Do you know other instrument types than the ones above to gain wind information?
- What are the measurement principles?
- What are the advantages/disadvantages of the different sensor types?

5 Soil Measurements

You will learn about soil moisture measurements in the corresponding experiment. This is an overview of the actual instrumentation:



(1,2) Soil moisture (University of Berne 1502B) and soil temperature (ETHZ Pt100, Campbell T107) both at 7 levels
(3) Soil moisture (IMKO TRIME-IT/EZ, Decagon 10HS, Campbell CS616, FZK SiSo) and soil temperature (Campbell T107) at 7 levels, soil heat flux (Hukseflux HFP01), tensiometer (UMS T8), and soil temperature (Hukseflux STP01). Not on the picture is the COSMOS probe to measure soil moisture.

Theory about soil heat flux G derived from soil temperature:

The rate at which heat flows through a soil at a depth z below the surface is directly proportional to the temperature gradient:

$$G = -\lambda \frac{\partial T}{\partial z}$$

where λ is the thermal conductivity ("Wärmeleitfähigkeit") [W m^{-1} K^{-1}].

The thermal conductivity λ is the ability of the soil to transport thermal energy. It corresponds to the energy (J), which passes vertically a horizontal area of 1 m^2 within one second if the temperature gradient is 1 K m^{-1}. λ depends strongly on soil moisture content.

Changes of G with depth lead to changes in of the heat content of the soil over time period Δt:

$$\frac{\partial T}{\partial t} = \frac{-1}{C} \frac{\partial G}{\partial z}$$

where C is the heat capacity ("Wärmekapazität") [J m^{-3} K^{-1}]. It is the energy (J) used to warm up 1 m^3 of soil by 1 K.

Theory about soil heat flux G derived from soil heat flux plates:

In order to calculate the soil heat flux density at the soil surface G the values of the heat flux plates G_z, which are buried at a given depth z (here -0.05 m), have to be adjusted for the changes in the heat storage in the layer between the soil surface and the sensor Δz. This additional term depends on the temperature change $\Delta(T_{soil})$ at the same depth z over time period Δt and the soil heat capacity c_v (Fuchs and Tanner 1968):

$$G = G_z + c_v \frac{\Delta T_{soil}}{\Delta t} \Delta z$$

The average heat capacity of the soil is given by its relative content of minerals x_{min}, organic matter x_{org}, water x_w and air x_p and their typical values of heat capacity, 2.1, 2.5, 4.2, and 0.0013 MJ m^{-3} K^{-1}, respectively (van Wijk and de Vries, 1963; Scheffer et al., 1998):

$$c_v = c_{min} x_{min} + c_{org} x_{org} + c_w x_w + c_p x_p$$

The content of organic matter can be assumed to be 2 % and the one of minerals 40 %. The volumetric content of water is measured and the remainder is assumed to be air.

Questions:

- What is the measurement principle of a heat flux plate?
- What are its limitations?

6 Discharge

You will hear more about discharge/runoff measurements. Here the three gauging sites with three different weir types:

Gauge oberer Rietholzbach (OTT ODS4)

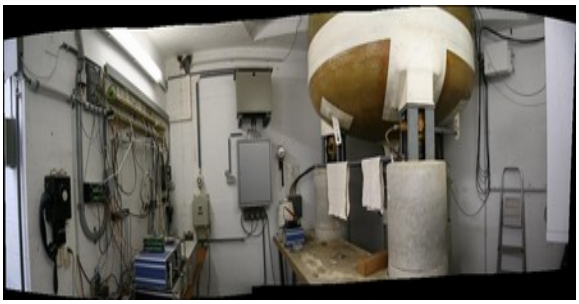
Gauge Huwilerbach (OTT PLS)

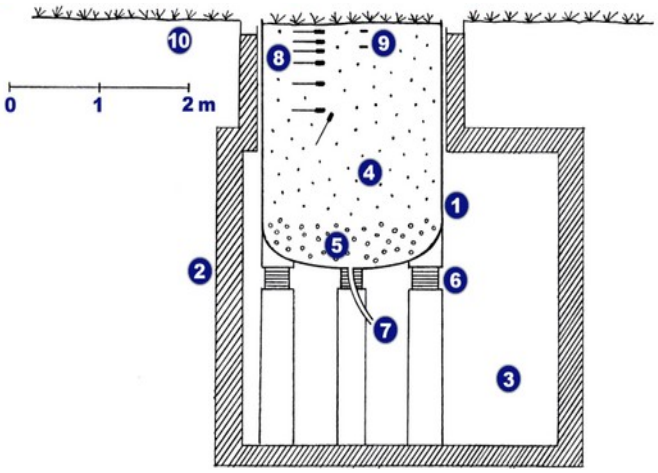
Gauge Rietholz-Mosnang

7 Evaporation (and Precipitation)

Lysimeter

Top and bottom view of the lysimeter in the upper row and a schematic in the lower row.





1 Container
2 Concrete wall
3 Cellar
4 Soil (gleyic cambisol)
5 Filter (sand and gravel)
6 Electronic balance
7 Drainage outlet
8 Soil moisture sensors (TDR)
9 Soil temperature sensors
10 Grass

It is rather difficult to measure actual evapotranspiration. A lysimeter is a very stationary and costly way.

- How can one derive evapotranspiration from lysimeter values? What are the necessary assumptions?
- Where do problems arise from the setting of such a lysimeter?
- Do you have other ideas to measure evapotranspiration?

a) $E = \boxed{P} - \frac{W_{t+1} - W_t}{\rho_w A} - Q$

This is the equation for calculating evapotranspiration mentioned in the movie.

8 Air Temperature and Air Humidity

Air temperature and humidity are measured with different sensors/techniques at Büel.

(Meteolabor Thygan VTP6)

(Rotronic HC2-S3)

(Campbell CSAT3)

- What are the different measurement techniques?
- What are the advantages/disadvantages?

9 Turbulent Fluxes (incl. Evaporation)

By the combination of an ultrasonic anemometer thermometer (short-form: sonic) and an open-path CO_2/H_2O infrared gas analyzer (short-form: IRGA) a couple of variables can be derived:

- Wind vector
- Air temperature
- Momentum flux
- Sensible heat flux
- CO_2 concentration/flux
- Water vapour concentration
- Latent heat flux/Evaporation

Turbulence measurements (wind vector, temperature, CO_2/H_2O concentration)(Campbell CSAT3, LiCor Li7500)

Operation principle of a sonic (anemometer):

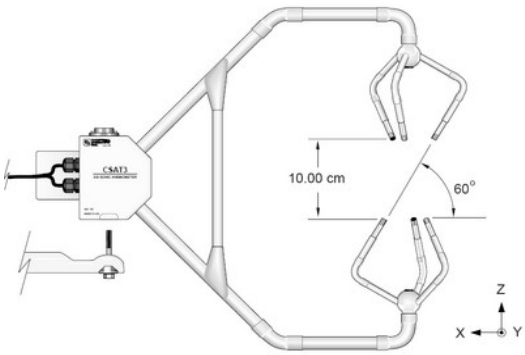
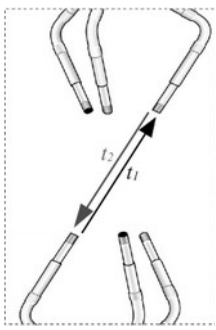


FIGURE 1. CSAT3 Coordinate System

(www.campbellsci.com)



Time of flight of the signal out

$$t_1 = \frac{d}{c + u_a}$$

Time of flight of the signal back:

$$t_2 = \frac{d}{c - u_a}$$

Wind speed along the transducer axis:

$$u_a = \frac{d}{2}\left[\frac{1}{t_1} - \frac{1}{t_2}\right]$$

Speed of sound along the transducer axis:

$$c = \frac{d}{2}\left[\frac{1}{t_1} + \frac{1}{t_2}\right]$$

Where d is the distance between the transducers and c is the speed of sound.

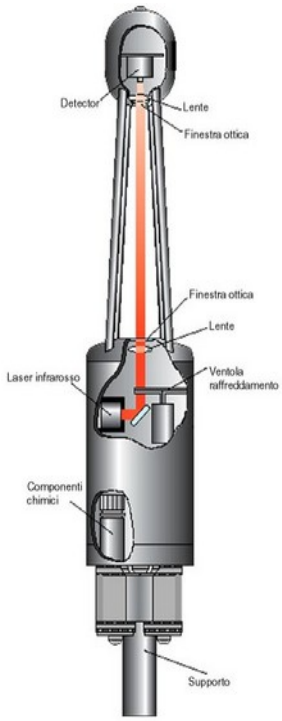
The wind speed is measured along all three non-orthogonal axis. These components are then transformed into an orthogonal coordinate system, which results in the wind vector.

As the speed of sound c depends on air temperature T and humidity q the air temperature can be derived as well from:

$c^2 = \gamma R_d T (1 + 0.61 q)$,

where γ is the ratio of specific heat of moist air at constant pressure c_p to that at constant volume c_v, and R_d is the gas constant for dry air.

Operation principle of an IRGA:



(basesperimentale.le.isac.cnr.it)

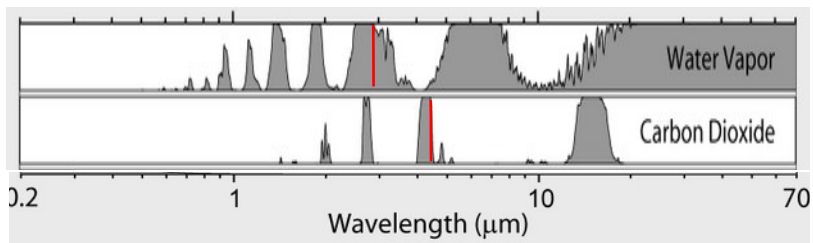
The absorption by a gas i is approximated by

$$\alpha_i = \left(1 - \frac{A_i}{A_{i0}}\right)$$

where A_i is the power received from the source in an absorbing wavelength and A_{i0} the power received from the source in a reference wavelength that does not absorb gas i.

The Li7500 measures A_i and A_{i0} alternately at 152 Hz.

The detector is a thermo-electrically cooled lead selenide, operating at wavelengths between 1.5 and 5.2 μm. The absorption at 2.59 μm and 4.26 μm results in the values for water vapour and CO_2 respectively. The cross-sensitivity is corrected internally.



Atmospheric absorption bands. (Modified after www.globalwarmingart.com)

Eddy covariance method

Turbulent flows, induced by shear stress and buoyancy, consist of many different size turbulence elements, the eddies. They transport physical properties such as momentum or CO_2. Thus, the vertical flux density at a given point in space can be determined as the product of the vertical wind component and the property of interest. As turbulence is highly variable and chaotic in space and time, it can be treated as a stochastic process. Hence, to get a reliable estimate of the vertical flux density an ensemble average should be calculated. In practice it is neither possible to make an average over many situations under identical conditions at one given point nor to carry out measurements at any point in a horizontal plane at a given height. Fortunately, the *ergodic hypothesis* can be made, i.e., spatial and time average converge over an

appropriate time interval to the ensemble average. Additionally, *Taylor hypothesis* of frozen turbulence allows time series measured at a single point to be interpreted as spatial variations, providing that the time series contains all information about the size distribution of the eddies.

Horizontal homogeneity simplifies the determination of vertical flux densities, because advective terms can be ignored. Hence, the statistical characteristics only vary in the vertical. Homogeneity is given if an adequate fetch is present and therefore the flow can be considered as adapted to the surface. If the turbulent characteristics do not vary with time the time series are *statistically stationary*. Under this condition *Reynolds decomposition* can be applied to separate the instantaneous value of a variable x in its mean value (denoted by an overbar) and its fluctuation from the mean (denoted by a prime):

$$x(t)=\overline{x}+x'(t)$$

Applying the ergodic hypothesis and the assumption of homogeneity the vertical flux density can be calculated as the covariance between the vertical wind component w and a property of interest x:

$$cov(w,x)=\frac{1}{N}\sum_{t=1}^{N}(w_t-\overline{w})(x_t-\overline{x})$$

The Reynolds averaging conditions simplify the calculation of the vertical flux density. They can be summarized as: (i) all fluctuating quantities average to zero, (ii) correlations between fluctuating and average quantities disappear.

Applying these assumptions and assuming that the average vertical wind component equals zero, the vertical flux density F becomes:

$$F=cov(w,x)=\frac{1}{N}\sum_{t=1}^{N}w_t'x_t'=\overline{w'x'}$$

Accordingly, the turbulent flux densities of sensible heat Q_H, latent heat Q_E, carbon dioxide F_c and momentum τ are calculated as:

$$Q_H=\rho c_p\overline{w'\theta'}$$

$$Q_E=l_v\overline{w'q'}$$

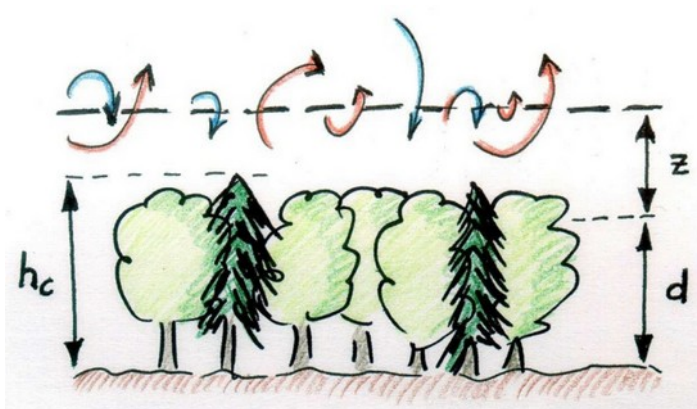
$$F_c=\overline{w'c'}$$

$$\tau=-\rho\overline{w'u'}$$

where ρ is the air density, c_p specific heat of moist air at constant pressure, and l_v the latent heat of vaporization.

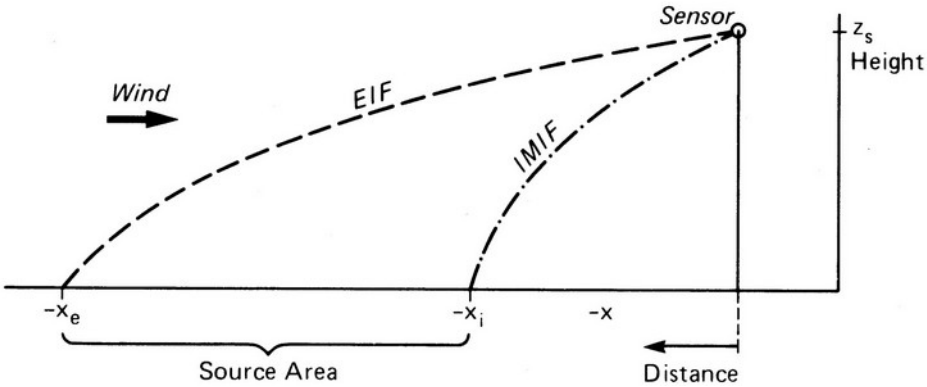
Measurement height

The following schematic (by W. Eugster) illustrates the concept of eddy covariance flux measurement height over tall vegetation. In this case, turbulence above the vegetation is a function of the distance above displacement height d. As a rule of thumb, d is 2/3 of the canopy height h_c if the canopy is homogeneous. Close to the surface, i.e., within the roughness sublayer, the flow is directly affected by individual roughness elements and the flow has to be treated as three dimensional. Thus, the flow is not in local equilibrium and local advection and horizontal turbulent transport processes are not negligible. The depth of the roughness sublayer is about twice the mean obstacle height depending on the size and the allocation of the roughness elements. As another rule of thumb, the *roughness length* of a canopy z_0 is about 0.1 h_c, and the roughness sublayer has the size of $z_* \sim 100\ z_0$. Thus, optimum measurement height is $z_m > z_*$. In reality this is not always possible, particularly over tall vegetation, and the problem arises that the point measurement is not automatically representative for the larger surface area.

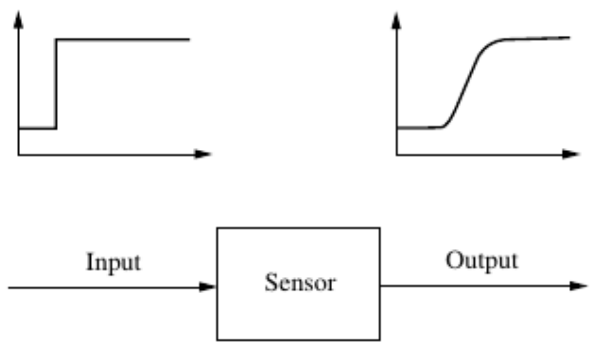


Source area

One important aspect is the representativeness of the measurements at a single tower for the real ecosystem fluxes. The measurements are influenced by surface elements within a given area in upwind distance, the *source area*. Here a schematic from Schmid and Oke (1990) of the source area concept is shown.

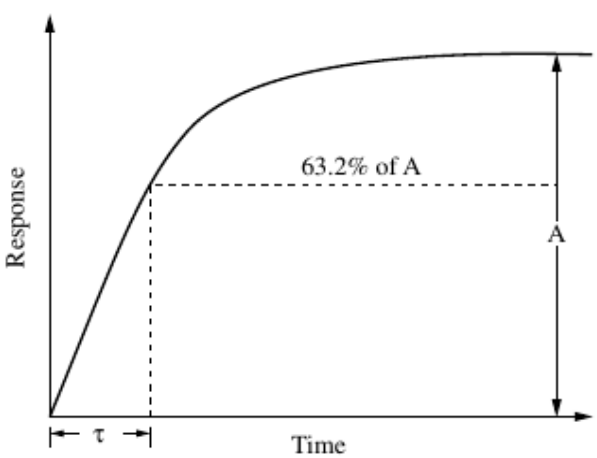


Time constant



The time constant (τ) is defined as the time required for the output of a sensor to complete 63.2% of the total rise (or decay) of an entity resulting from a step change in its value.

$\tau = 1 - 1/e = 0.632$



Lipták, B.G. (1999) Instruments engineer's handbook.

Questions:

- Which instruments/measurement principles are characterised by rather large/short time constants?
- How do you expect a long time constant to influence observational data?
- How can the influence of the time constant on observational data be reduced?

References and Further Readings

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