Surface Energy Balance

Radiation balance / Net radiation $NR$

Evapotranspiration
Latent heat flux $L_vE$

Sensible heat flux $H$

Soil heat flux $G$

Available Energy

Turbulent Fluxes
Surface Energy Balance

Top of the atmosphere

- Kinetic energy: 1.4 MJ m\(^{-2}\)
- Static energy: 2620 MJ m\(^{-2}\)
- Internal energy: 1880 MJ m\(^{-2}\)
- Potential energy: 740 MJ m\(^{-2}\) (APE: 5.3 MJ m\(^{-2}\))
- Net Radiation: +104 W m\(^{-2}\)

Surface

- Sensible heat flux: -104 W m\(^{-2}\)
- Latent heat flux: -85 W m\(^{-2}\)
- Global radiation: 169 W m\(^{-2}\)
- Reflected radiation: -25 W m\(^{-2}\)
- Longwave radiation downward: 345 W m\(^{-2}\)
- Longwave radiation upward: -385 W m\(^{-2}\)

Fig. 10.5 Redistribution of radiative energy in the climate system (from 98Wil and 04Ohm). All values refer to a mean solar constant of 1368 Wm\(^{-2}\) and are global annual averages. Note the large amount of surplus heat at the surface, which must be removed from the surface by turbulent fluxes of latent and sensible heat. The radiation budget of the atmosphere alone is negative, requiring an amount of about 104 Wm\(^{-2}\) to be added from ground by evapotranspiration and fluxes of sensible heat.

(Ohmura&Raschke 2005)
Surface Energy Balance

The surface energy balance is usually defined with respect to an active layer of infinitesimal small thickness. In this case the storage of energy in the layer can be neglected and the energy balance equation takes the form:

\[ S \downarrow (1 - \alpha) + L \downarrow - L \uparrow + H + L_v E + G = 0 \]

or, summarizing the radiation fluxes:

\[ NR + H + L_v E + G = 0 \]

where

- **NR**: net radiation
- **S\downarrow**: global radiation
- **\alpha**: albedo
- **L\downarrow**: atmospheric counter-radiation
- **L\uparrow**: terrestrial emission
- **H**: sensible heat flux
- **L_v E**: latent heat flux
- **G**: soil heat flux

**Sign convention**

Fluxes are considered **positive** when directed toward the surface (energy sources) and **negative** when directed away from the surface (energy sinks). Exceptions are \( L \uparrow \) and a \( S \uparrow \) (outgoing radiation fluxes), for which a minus sign is explicitly used in the energy balance equation.
Energy Balance of a Volume

In other situations, however, the active layer has a measurable thickness. In this case the rate of change of energy stored in the layer, \( \frac{\partial Q}{\partial t} \), must be included in the equation:

\[
NR + H + L_v E + G = \frac{\partial Q}{\partial t}
\]  

In many instances we also need to take the lateral fluxes (advection) into account. This situation is encountered for instance with vegetation or snow.

(Oke 1987)
Global Energy Balance

The first attempt to produce an atlas of the surface energy balance is due to Budyko (1956) and the results are summarized in Budyko (1974). The results presented here are taken from Raschke and Ohmura (2005) and Ohmura and Raschke (2005). They are based on satellite measurements from the International Satellite Clouds Climatology Project (ISCCP) and data from the Reanalysis Project of the European Centre for Medium-Range Weather Forecasts (ECMWF ReAnalysis, ERA-40).

References


Daily sum of incoming solar radiation at the top of the atmosphere (Lauer & Bendix 2004)
Radiation Balance 1: Global Radiation

Annual average of the global radiation at ground during the period 1991-1995 (Raschke&Ohmura 2005)
Radiation Balance 2: Albedo

\[ \alpha = \frac{S \uparrow}{S \downarrow} \]

- desert: 0.20 – 0.45
- grasland: 0.16 – 0.26
- forest: 0.15 – 0.20
- city: 0.10 – 0.20
- water: 0.03 – 0.10 (flat angle up to 1.00!)
- fresh snow: 0.95
- snow: 0.45 – 0.95

(Oke 1987)
Annual average of the albedo of the earth-atmosphere system during the period 1991-1995 (Raschke&Ohmura 2005)
Radiation Balance 3: Net shortwave

Annual average of the solar net radiation at ground during period 1991-1995 (Raschke&Ohmura 2005)
Annual average of the downward longwave radiation at ground during the period 1991-1995 (Raschke&Ohmura 2005)
Annual average of the terrestrial emission at top of the atmosphere during the period 1991-1995 (Raschke&Ohmura 2005)
Annual average of the net fluxes of longwave radiation during the period 1991-1995 (Raschke&Ohmura 2005)
Annual average of the net radiation at the surface during the period 1991-1995 (Raschke & Ohmura 2005)
Annual mean distribution of sensible heat fluxes during the period 1991-1995 (Raschke&Ohmura 2005)
Latent Heat Flux

$L_v E$

Annual mean distribution of latent heat fluxes during the period 1991-1995 (Raschke & Ohmura 2005)
Temporal Course Example 1

Grasland in Northeastern Switzerland (8.52°E, 47.43°N)
Temporal Course Example 2

Tropical rain forest in Brazil, State of Rondonia (61.93°W, 10.08°S)
Temporal Course Example 3

Arctic sea ice (drifting station, ~160°W, 75°N)
Data Archive GEBA – Global Energy Balance Archive

- GEBA is a project of the World Climate Programme – Water (WMO/ICSU)
- Data are collected since 1985 at ETH by Ohmura, Wild and co-workers
- monthly surface values
- more than 2000 Stationen worldwide
- http://proto-gebra.ethz.ch/

(Ohmura&Raschke 2005)
Soil Heat Flux 1

\( G \)

(Lecture A. Christen 2004)
Soil Heat Flux 2

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

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

$\lambda$ is the thermal conductivity (‘Wärmeleitfähigkeit’) [W m$^{-1}$ K$^{-1}$].

(Lecture A. Christen 2004)
The thermal conductivity $\lambda$ is the ability of the soil to transport thermal energy. It corresponds to the energy (J), which passes vertically a horizontal area of m$^2$ within one second if the temperature gradient is 1 K m$^{-1}$.

<table>
<thead>
<tr>
<th>Material</th>
<th>State</th>
<th>Thermal conductivity $\lambda$ (W m$^{-1}$ K$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy Soil</td>
<td>Dry</td>
<td>0.30</td>
</tr>
<tr>
<td></td>
<td>(40% porosity)</td>
<td>2.20</td>
</tr>
<tr>
<td>Clay Soil</td>
<td>Dry</td>
<td>0.25</td>
</tr>
<tr>
<td></td>
<td>(40% porosity)</td>
<td>1.58</td>
</tr>
<tr>
<td>Marshland</td>
<td>Dry</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td>(80% porosity)</td>
<td>0.50</td>
</tr>
<tr>
<td>Snow</td>
<td>Old</td>
<td>0.08</td>
</tr>
<tr>
<td></td>
<td>Fresh</td>
<td>0.42</td>
</tr>
<tr>
<td>Ice</td>
<td>0 °C, pure</td>
<td>2.24</td>
</tr>
<tr>
<td>Water</td>
<td>4 °C</td>
<td>0.57</td>
</tr>
<tr>
<td>Concret</td>
<td></td>
<td>1.51</td>
</tr>
</tbody>
</table>

(Lecture A. Christen 2004)
Soil Heat Flux 4

Changes of $G$ with depth lead to changes in time $t$ of the heat content of the soil:

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

$C$ is the heat capacity (‘Wärmekapazität’) [J m$^{-3}$ K$^{-1}$].

$C$ is the energy (J) used to warm up 1 m$^3$ of soil by 1 K.

<table>
<thead>
<tr>
<th>Material</th>
<th>State</th>
<th>Heat Capacity C (J m$^{-3}$ K$^{-1}$ × 10$^6$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy soil</td>
<td>Dry</td>
<td>1.28</td>
</tr>
<tr>
<td>(40% porosity)</td>
<td>Saturated</td>
<td>2.96</td>
</tr>
<tr>
<td>Clay Soil</td>
<td>Dry</td>
<td>1.42</td>
</tr>
<tr>
<td>(40% porosity)</td>
<td>Saturated</td>
<td>3.10</td>
</tr>
<tr>
<td>Marshland</td>
<td>Dry</td>
<td>0.58</td>
</tr>
<tr>
<td>(80% porosity)</td>
<td>Saturated</td>
<td>4.02</td>
</tr>
<tr>
<td>Snow</td>
<td>Fresh</td>
<td>0.21</td>
</tr>
<tr>
<td></td>
<td>Old</td>
<td>0.84</td>
</tr>
<tr>
<td>Ice</td>
<td>0 °C, pure</td>
<td>1.93</td>
</tr>
<tr>
<td>Water</td>
<td>4 °C</td>
<td>4.18</td>
</tr>
<tr>
<td>Concret</td>
<td></td>
<td>2.11</td>
</tr>
</tbody>
</table>

(Lecture A.Christen 2004)
Combining the two equations and assuming that the thermal conductivity does not vary with depth, we arrive at a second-order partial differential equation for the soil temperature with the following solution:

\[
\frac{\partial T}{\partial t} = -\frac{1}{C} \frac{\partial}{\partial z}\left(-\lambda \frac{\partial T}{\partial z}\right) = \frac{\lambda}{C} \frac{\partial^2 T}{\partial z^2} \equiv \kappa \frac{\partial^2 T}{\partial z^2}
\]

\[
T(z, t) = \langle T \rangle + \Delta T_0 e^{-z\sqrt{\omega/2\kappa}} \sin(\omega t - z\sqrt{\omega/2\kappa})
\]

\(\kappa = \lambda/C\) is the thermal diffusivity (‘Temperaturleitfähigkeit’) \([\text{m}^2 \text{s}^{-1}]\).

\(\kappa\) describes the 'travel' velocity of a temperature change in the soil and controls the penetration depth of changes in temperature at the surface.

<table>
<thead>
<tr>
<th>Material</th>
<th>State</th>
<th>Thermal diffusivity (\kappa) ([\text{m}^2 \text{s}^{-1} \times 10^{-6}])</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy soil</td>
<td>Dry</td>
<td>0.24</td>
</tr>
<tr>
<td>(40% porosity)</td>
<td>Saturated</td>
<td>0.74</td>
</tr>
<tr>
<td>Snow</td>
<td>Fresh</td>
<td>0.10</td>
</tr>
<tr>
<td></td>
<td>Old</td>
<td>0.40</td>
</tr>
<tr>
<td>Water</td>
<td>4 °C</td>
<td>0.24</td>
</tr>
<tr>
<td>Concrete</td>
<td></td>
<td>0.72</td>
</tr>
</tbody>
</table>
Surface Processes

The energy exchange is only but one of the possible links between the hydrosphere/biosphere/cryosphere/pedosphere and the atmosphere. Other interactions include:

- the exchange of water;
- the exchange of trace constituents (CO₂, CH₄, N₂O, ...);
- the exchange of momentum (friction!)
Atmospheric Boundary Layer 1

If we actually look closer at these interactions, we see that they involve not only the very surface but also the atmospheric layer close to the surface. This is the region called the atmospheric or planetary boundary layer (ABL or PBL for short).

Further Reading
Atmospheric Boundary Layer 2

The atmospheric boundary layer can be defined as that part of the troposphere that is directly influenced by the presence of the earth’s surface, and responds to surface forcings with a timescale of about an hour or less (Stull 1988).

Fig. 3.1. Definition sketch showing orders of magnitude of the heights of the sublayers of the atmospheric boundary layer (ABL); $h_r$ is a typical height of the roughness obstacles; the (distorted) vertical scale is in meters.
Atmospheric Boundary Layer 3

The ABL has a well defined structure that evolves with the diurnal cycle:

Fig. 1.7 The boundary layer in high pressure regions over land consists of three major parts: a very turbulent mixed layer; a less-turbulent residual layer containing former mixed-layer air; and a nocturnal stable boundary layer of sporadic turbulence. The mixed layer can be subdivided into a cloud layer and a subcloud layer. Time markers indicated by S1-S6 will be used in Fig. 1.12.
Atmospheric Boundary Layer 3 (in colour)

Free Atmosphere

E.Z. Capping Inversion

Mixed Layer Residual Layer

Stable BL

Height, Z

Day 1 Night 1 Day 2


(Quelle: Wallace&Hobbs 2006)
The top of the ABL is usually detectable in soundings of the lower atmosphere. The following figures show examples of profiles of the potential temperature, specific humidity and wind direction measured in the upper Ticino on August 25, 1999, 12:00 UTC. They refer to a convective situation. Idealized profiles are superimposed in red.
One of the essential features of the flow in the ABL is that it is turbulent. Turbulence can be characterized as follows:

- turbulence is a random (stochastic), 3-dimensional, rotational motion;
- turbulence is non-linear;
- turbulence can only be treated statistically;
- turbulence is diffusive. It efficiently transports momentum, heat, water vapor and other constituents (e.g. CO₂);
- turbulence is dissipative. Molecular viscosity is responsible for the decay of kinetic energy into heat (increase in internal energy at the expense of mechanical energy). To exist turbulence must continuously feed on the ‘mean’ flow.
Reynolds Number 1

Laminar flow: 
Transport by 
Molecular 
diffusion

Turbulent flow: 
$10^6$ times more efficient 
than laminar flow

(Wallace&Hobbs 2006)
Reynolds Number 2

A distinction between laminar and turbulent flows is possible on the basis of the Reynolds number $Re$ (Batchelor, 1967). For $Re > 2000$, small perturbations of the flow rapidly grow, eventually leading to turbulence.

The Reynolds number is defined as the ratio of the inertia and viscous term in the equation of motion. Assuming stationarity:

$$u_k \frac{\partial u_i}{\partial x_k} = -\rho^{-1} \frac{\partial p}{\partial x_i} + \nu \frac{\partial^2 u_i}{\partial x_k \partial x_k}$$

where $\nu \sim 10^{-5}$ m$^2$ s$^{-1}$ is the kinematic viscosity. In orders of magnitude, given velocity and length scales $U$ and $L$:

$$\frac{U^2}{L} \sim -\rho^{-1} \frac{P}{L} + \nu \frac{U}{L^2}$$

whereby, upon building the ratio of the first and last term, one obtains:

$$Re \equiv \frac{U^2 / L}{\nu U / L^2} = \frac{U L}{\nu}$$

Reynolds Number 3

In the atmospheric boundary layer $U \sim 1$ to $10 \text{ m s}^{-1}$ (a typical wind speed) and $L \sim 1$ to $1000 \text{ m}$ (the height above ground). Hence typical values for the Reynolds number are:

$$\text{Re} = \frac{UL}{\nu} \sim \frac{1 \cdot 1}{10^{-5}} \text{ to } \frac{10 \cdot 1000}{10^{-5}} = 10^5 \text{ to } 10^7 \gg 2000$$

It follows that with the exception of a very thin layer close to the surface, ABL flows are always turbulent.

(Quelle: Oke 1987)
Eddy Covariance Method 1

A succession of correlated perturbations in a time series eventually lead to a statistical correlation over some finite interval of time:

Figure 2.12 The relationships between vertical velocity (w) and air temperature (T) fluctuations, and the instantaneous sensible heat flux ($Q_H$). Results from fast-response instruments at a height of 23 m over grass in unstable conditions (after Priestley, 1959).

(Quelle: Oke 1987)
Eddy Covariance Method 2

In looking at turbulence it is convenient to treat it as a perturbation \((u'_i, \theta', q')\) for the wind components, potential temperature and specific humidity, respectively) of a ‘mean’ flow \((U_i, \Theta, Q)\). That is to say we introduce the so-called Reynolds decomposition:

\[
\begin{align*}
    u_i &= U_i + u'_i, \quad i=1,2,3 \\
    \theta &= \Theta + \theta' \\
    q &= Q + q'
\end{align*}
\]

(Quelle: Vorlesung R.Vogt 2006)
Eddy Covariance Method 3

Turbulent transfer arises from the covariance between the fluctuating components of the flow. With respect to the transfer of momentum \( \tau \), sensible heat \( H \), and latent heat \( L_v E \), and using the notation \((u, v, w) \equiv (u_1, u_2, u_3)\) for the three components of the wind vector, we have:

\[
\begin{align*}
\tau & \equiv -\rho w'u' \\
H & \equiv \rho C_p \overline{w'\theta'} \\
L_v E & \equiv \rho L_v \overline{w'q'}
\end{align*}
\]

Sign convention
In boundary layer studies, the shear stress \( \tau \) is by definition positive when there is an associated transport of momentum to the surface, but \( H \) and \( L_v E \) are positive when directed away from the surface.
Correlation of Turbulent Components

How do we explain the correlation? Consider the following ‘mean’ profiles for the longitudinal wind component \( U(z) \) and the potential temperature \( \theta(z) \), the associated shear stress \( \tau > 0 \) and heat transfer \( H > 0 \), and the perturbations \( u' \) and \( \theta' \) induced by a vertical displacement \( w' > 0 \). We have:

\[
\begin{align*}
\tau > 0, \ w'u' & < 0 \\
\theta' > 0, \ w'\theta' & > 0
\end{align*}
\]
Turbulent Fluxes and Mean Gradients 1

Fluctuating components and their correlation can be directly measured. However, this type of data is not always available. For this reason, ways for relating the turbulent fluxes to the vertical variations of the mean fields are sought.

There are two basic approaches. In the first, turbulent fluxes are considered in the same way as molecular fluxes. Then:

\[ \tau \equiv -\rho \overline{w' u'} = \rho K_M \frac{\partial U}{\partial z} \]

\[ H \equiv \rho C_p \overline{w' \theta'} = -\rho C_p K_H \frac{\partial \Theta}{\partial z} \]

\[ L_v E \equiv \rho L_v \overline{w' q'} = -\rho L_v K_E \frac{\partial Q}{\partial z} \]

where \( K_M \) is the eddy (turbulent) viscosity and \( K_H \) and \( K_E \) are the eddy diffusivities. All \( K \)'s (the ‘Austauschkoefizienten’) are in units of \([\text{m}^2 \text{ s}^{-1}]\).
The second approach makes use of the description of an electrical current $I$ in terms of an electric potential $V$ and a resistance $R$:

$$I = \frac{V}{R}$$

If this concept is applied to the turbulent transfer, we obtain:

$$\tau \equiv -\rho w'u' = \rho \frac{U - U_0}{r_{aM}} = \rho \frac{U}{r_{aM}}$$

$$H \equiv \rho C_p w'\theta' = -\rho C_p \frac{\Theta - \Theta_0}{r_{aH}}$$

$$L_v E \equiv \rho L_v w'q' = -\rho L_v \frac{Q - Q_0}{r_{aE}}$$

where subscript 0 denotes surface values, and where $r_{aM}$, $r_{aH}$ and $r_{aE}$ are aerodynamic resistances in units of [m$^{-1}$ s].

A problem with this approach is the fact that the definition of ‘surface’ values is not always simple.
A General Transfer Scheme

The description of a ‘transfer’ in terms of a potential and a resistance can be extended to all scales involved in the interactions between the surface and the atmosphere. Resistances can be put in series or parallel, depending on the format of the transfer. For instance, for the flow of water from the soil, through vegetation, to the atmosphere we can apply the following scheme:

(Oke 1987)
Bowen Ratio

Experimental evidence indicates that aerodynamic transfer of heat and moisture through turbulence takes place in complete analogy. In general we assume that to first order $K_M \approx K_H \approx K_E$ and $r_{aM} \approx r_{aH} \approx r_{aE}$. This allows the Bowen ratio, $Bo \equiv H/L_v E$, to be expressed as:

$$Bo \equiv \frac{\rho C_p w' \theta'}{\rho L_v w' q'} \approx \frac{C_p}{L_v} \frac{\partial \Theta}{\partial z} = \frac{C_p}{L_v} \frac{\Delta \Theta}{\Delta Q} \approx \frac{C_p}{L_v} \frac{(\Theta - \Theta_0)}{(Q - Q_0)}$$
Turbulent Transfer and the Stability of the Atmosphere

Depending on the vertical variation of the potential temperature, the atmosphere can be statically stable, neutral or unstable with respect to adiabatic displacements. We have:

\[
\frac{\partial \Theta}{\partial z} > 0 \quad \Rightarrow \quad \text{stable stratification}
\]

\[
\frac{\partial \Theta}{\partial z} = 0 \quad \Rightarrow \quad \text{neutral stratification}
\]

\[
\frac{\partial \Theta}{\partial z} < 0 \quad \Rightarrow \quad \text{unstable stratification}
\]

Under stable conditions, vertical displacements are limited by buoyancy because denser/lighter fluid finds itself in a lighter/denser environment. This means that also the vertical motion induced by turbulence is reduced or enhanced depending on the sign of \( \frac{\partial \Theta}{\partial z} \).

To account for the effects of buoyancy on vertical displacements one introduces a stability correction either in the eddy diffusivities or the aerodynamic resistances.
Turbulent Transfer and the Stability of the Atmosphere

(Wallace & Hobbs 2006)
Richardson Number

On the other hand, turbulence is favored by high wind speeds, which means enhanced friction and mechanical production of instabilities by the wind shear, $\partial U/\partial z$. Therefore, the stability correction applied to the eddy diffusivities or resistances has to be further modified to account for the magnitude of the shear.

There are complementary ways to do so. A widely used possibility is to introduce the so-called **Richardson number** $R_i$, which is defined as:

$$R_i \equiv \frac{g}{\Theta_{ref}} \frac{\partial \Theta/\partial z}{(\partial U/\partial z)^2}$$

The denominator being always positive, the combined effects of stability and shear on turbulent transfer can be expressed as:

- $R_i > 0 \Rightarrow$ stable conditions
- $R_i = 0 \Rightarrow$ neutral conditions
- $R_i < 0 \Rightarrow$ unstable conditions
The Logarithmic Wind Profile

Two important parameters needed to describe the aerodynamic state of the flow are the friction velocity $u_*$ and the roughness length $z_0$. The friction velocity is defined as:

$$u_* \equiv \sqrt{\frac{\tau}{\rho}}$$

whereas the roughness length is the height at which $U = 0$. It is related to the geometric properties of the surface.

Together they characterize the shape of the so-called logarithmic wind profile, the vertical variation of wind speed under neutral conditions:

$$U(z) = \frac{u_*}{k} \ln \left( \frac{z}{z_0} \right)$$

where $k = 0.4$ is the von-Kármán constant.
Explicit Expression for $K_M$

Given the wind profile and the definition of the friction velocity, we can find explicit expressions for the eddy diffusivities and the aerodynamic resistances. We also need the first derivative of the wind profile, which in neutral conditions reads:

$$U(z) = \frac{u_*}{k} \ln \left( \frac{z}{z_0} \right) \Rightarrow \frac{\partial U}{\partial z} = \frac{u_*}{k z}$$

Then:

$$\tau = -\rho K_M \frac{\partial U}{\partial z} = -\rho K_M \frac{u_*}{k z} = \rho u_*^2$$

and by equating the last two terms:

$$K_M = k z u_*$$

To account for stability we usually write:

$$K_M = \frac{k z u_*}{\Phi(Ri)}$$

where $\Phi$ is a known function of the Richardson number.
Explicit Expression for $r_{aM}$

To arrive at explicit expressions for the resistances we introduce the wind profile directly into the definition of the shear stress:

$$\tau = \rho u_*^2 = \rho \frac{k^2 U^2}{\left(\ln\left(\frac{z}{z_0}\right)\right)^2} = \rho \frac{U}{r_{aM}}$$

showing that either:

$$r_{aM} = \frac{U}{u_*^2}$$

or, in neutral conditions:

$$r_{aM} = \frac{\left(\ln\left(\frac{z}{z_0}\right)\right)^2}{k^2 U}$$

To account for stability we usually write:

$$r_{aM} = \frac{\left(\ln\left(\frac{z}{z_0}\right) - \Psi(Ri)\right)^2}{k^2 U}$$

where $\Psi$ is another known function of the Richardson number.
In-situ measurement: Radiation

Precision tracker, automatic detection of sun’s position

IACETH, Greenland (2006)

IACETH, Rietholzbach (2009)
**In-situ measurement: Radiation**

Cavity radiometer (direct solar radiation). Measures in the wavelength range 335 to 2200 nm (0.33 to 2.2 mm).

Pyranometer (solar radiation). Schott K5 dome, 335-2200 nm.

Precision Infrared Radiometer (PIR) (longwave radiation). Si-dome + interference filter. Measures in the wavelength range ~ 4 to 50 mm.

*IACETH, Greenland (2006)*
In-situ measurement: Profiles and Eddy Covariance

CARBOEUROPE (2006)

IACETH, Rietholzbach (2009)
In-situ measurement: Mean values

Temperature and air humidity (aspirated instruments)

Wind speed and direction

Turbulence

IACETH, Greenland (2006)
In-situ measurement: Turbulence

Ultrasonic anemometer. Measures wind components AND temperature with a sampling rate of 20 to 50 s\(^{-1}\)

Ultrasonic anemometer + fast-response H\(_2\)O and CO\(_2\) sensors

* IACETH, Greenland (2006)
In-situ measurement: Temperature profile

IACETH, Greenland (2006)

Thermocouples. Use for detailed observations of the temperature profile.
In-situ measurement: ABL, free atmosphere

Radiosounding system. Use to measure profiles of horizontal wind vector, temperature, pressure and humidity up to ~ 25 km.

IACETH, Greenland (2006)