광합성과 증발산의 미기상학적 측정기술

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연세대학교

Technique for Estimating CO_2 and H_2O Exchange between the Atmosphere and the Biosphere : Eddy Covariance Method

Wonsik Kim
Yonsei University

◈ What is the flux ?

Flux is the transfer of a quantity per unit area per unit time.
The quantities are mass, heat, moisture, momentum and pollutant in micrometeorology.
Kinematic flux (Fluid)

$\overline{w}\cdot\overline{q}$

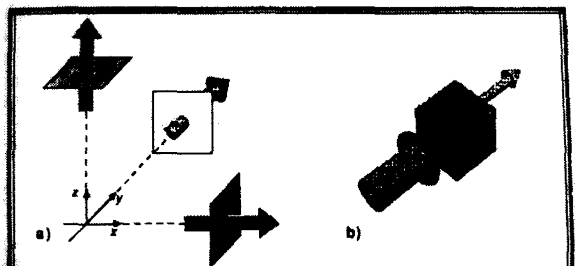


Fig. 2.10 Flux is the rate of transfer of a quantity across a unit area. (a) The net flux can be split into components in the x,y, and z directions, as shown here. (b) If a greater flux enters a volume than leaves, then there must be a net increase of that quantity within the volume.

Eddy flux (Turbulence)

$\overline{w'\cdot q'}$

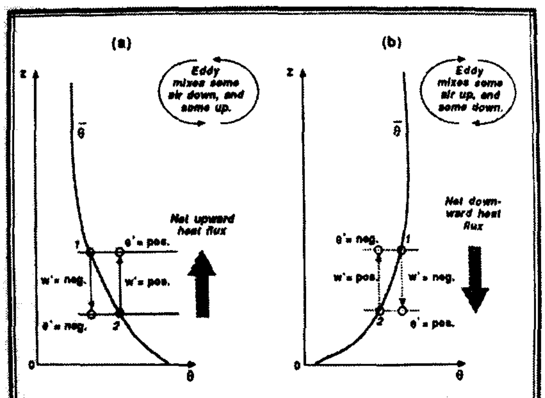


Fig. 2.12 Idealization of the small eddy mixing process, showing (a) net upward turbulent heat flux in a statically unstable environment, and (b) net downward turbulent heat flux in a stable environment.

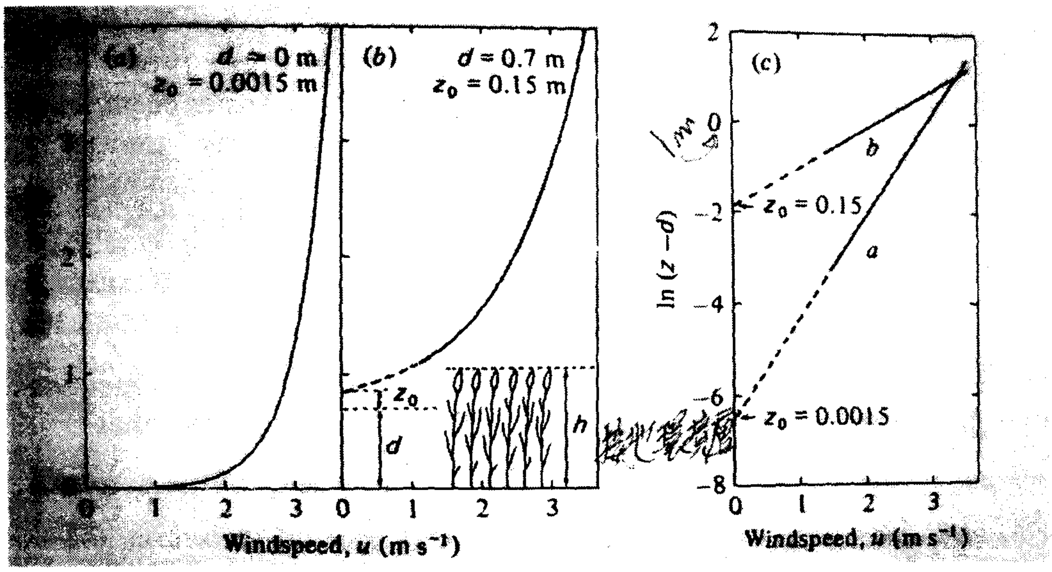
The reduction in wind speed due to frictional drag transfers momentum from the atmosphere to the surface, creating turbulence that mixes the air and transports heat and water from the surface into the lower atmosphere.

With greater height above the surface, eddies are larger so that transport of momentum, heat, and moisture is more efficient with height above the surface.

◈ Why we measure the flux ?

◇ **Momentum** (kg m s^{-1} / m^2 s)

To know the characteristics of boundary layers: behaviour of roughness length (z_0)and zero plane displacement (d), aerodynamic resistance



$$\tau = \rho \frac{u_a - u_s}{r_{am}}$$

$$\tau = \rho \left[\frac{k u_* (z-d)}{\phi_m(\zeta)} \right] \frac{\partial u}{\partial z}$$

$$u_2 - u_1 = \frac{u_*}{k} \left[\ln\left(\frac{z_2 - d}{z_1 - d}\right) - \psi_m\left(\frac{z_2 - d}{L}\right) + \psi_m\left(\frac{z_1 - d}{L}\right) \right]$$

$$u_z = \frac{u_*}{k} \left[\ln\left(\frac{z - d}{z_{0m}}\right) - \psi_k(\zeta) \right]$$

◇ **Energy** (J / m^2 s)

To know the heat budget at the surface and effects of the surface turbulent flux to atmosphere

$$R_n = (1-r)S\downarrow + (L\downarrow - L\uparrow) = H + \lambda E + G$$

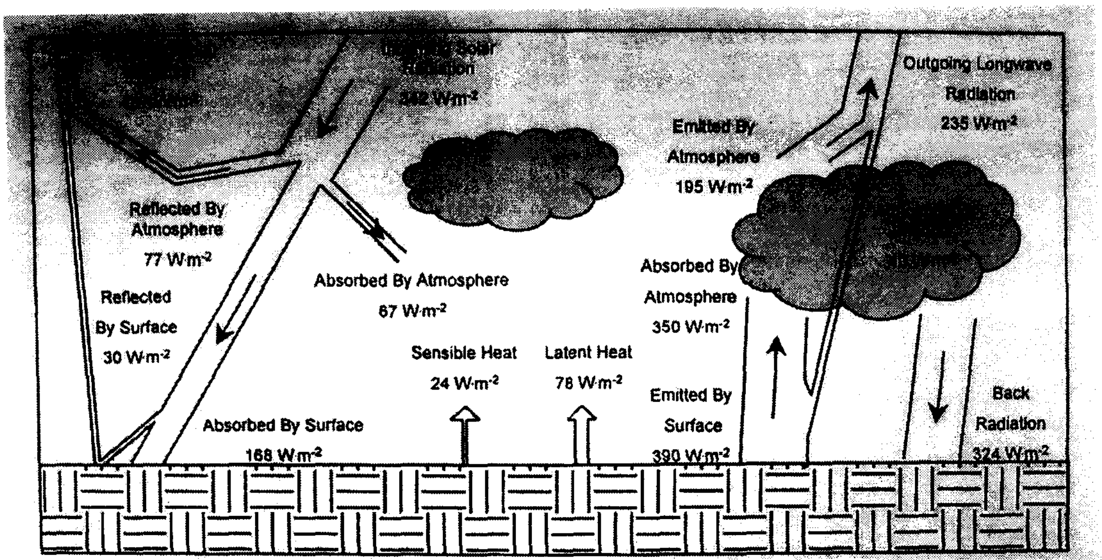


Figure 2.1. Earth's annual mean global energy budget showing solar radiation (left), sensible and latent heat fluxes (middle), and longwave radiation (right). Data from Kiehl and Trenberth (1997).

$$H = -\rho C_p \frac{T_a - T_s}{r_{ah}}$$

$$E = -\rho \frac{q_a - q_s}{r_{aw}}$$

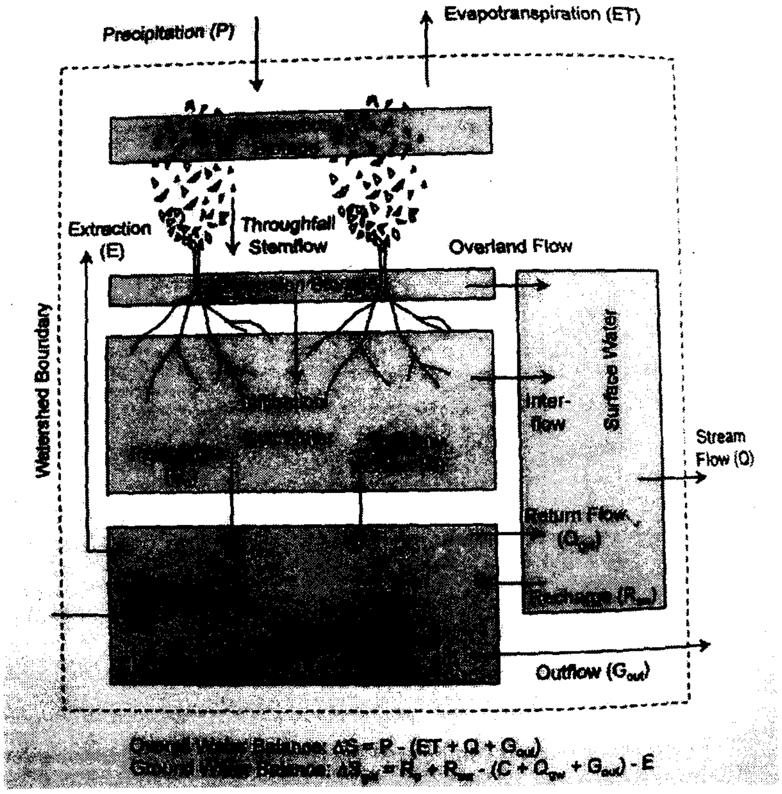
Penman-Monteith equation

$$\lambda E = \frac{s(R_n - G) + \frac{\rho C_p}{r_a}(e_a^* - e_a)}{s + \gamma\left(\frac{r_c - r_a}{r_a}\right)} = \frac{s(R_n - G) + \rho C_p g_a (e_a^* - e_a)}{s + \gamma\left(1 + \frac{g_a}{g_c}\right)}$$

◇ **Water** (kg / m^2 s)

To know the hydrologic cycle on land surface

$P = ET + Q$



◇ **Carbon dioxide** (kg / m^2 s)

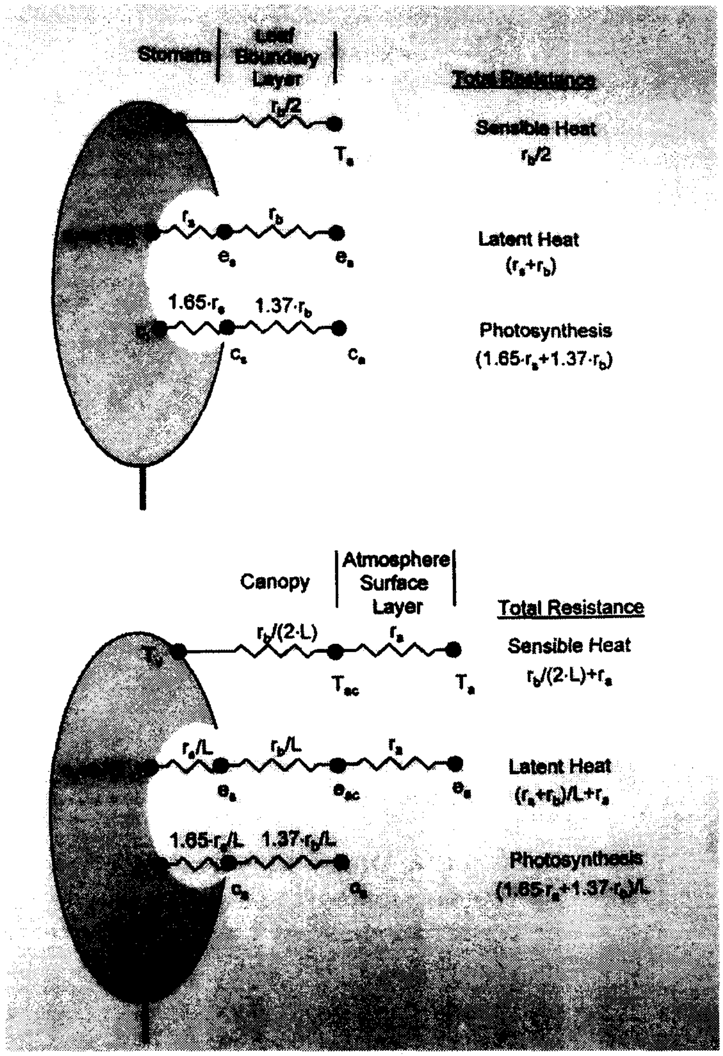
To understand the carbon dioxide budget above the ecosystem

$NEE = GPP - R$

$R = Rp + Rm$

◇ **Resistance** (s / m)

Flux density (or flux) = proportionality constant * driving force
Current = voltage / resistance



stomatal resistance:
Biological resistance due to stomatal movement controled photosynthetic activity and water status of plant.

boundary layer resistance:

$$r_b = \frac{\rho C_p d}{kNu} \approx 200\sqrt{\frac{d}{u}}$$

aerodynamic resistance:

$$r_{am} = \frac{1}{k^2 u_a}\left[\ln\left(\frac{z-d}{z_{0m}}\right) - \psi_m(\zeta)\right]\left[\ln\left(\frac{z-d}{z_{0m}}\right) - \psi_m(\zeta)\right]$$

canopy resistance:

$$r_{cH} = \frac{r_b}{2L}$$
$$r_{cW} = \frac{r_s + r_b}{L}$$

◈ How to measure the flux ?

◇ **Eddy covariance method**

Micrometeorological techniques have many advantage: First, they are in situ and do not disturb the environment around the plant canopy. Second, these techniques allow continuous measurements. And third, time-averaged micrometeorological measurements at a point provide an area-integrated, ensemble average of the exchange rates between the surface and the atmosphere.

Eddy covariance technique ascertains the exchange rate of CO_2 across the interface between the atmosphere and a plant canopy by measuring the covariance between fluctuations in vertical wind velocity and CO_2 mixing ratio.

General criteria for micrometeorological measurement are flat, steady and extended distance

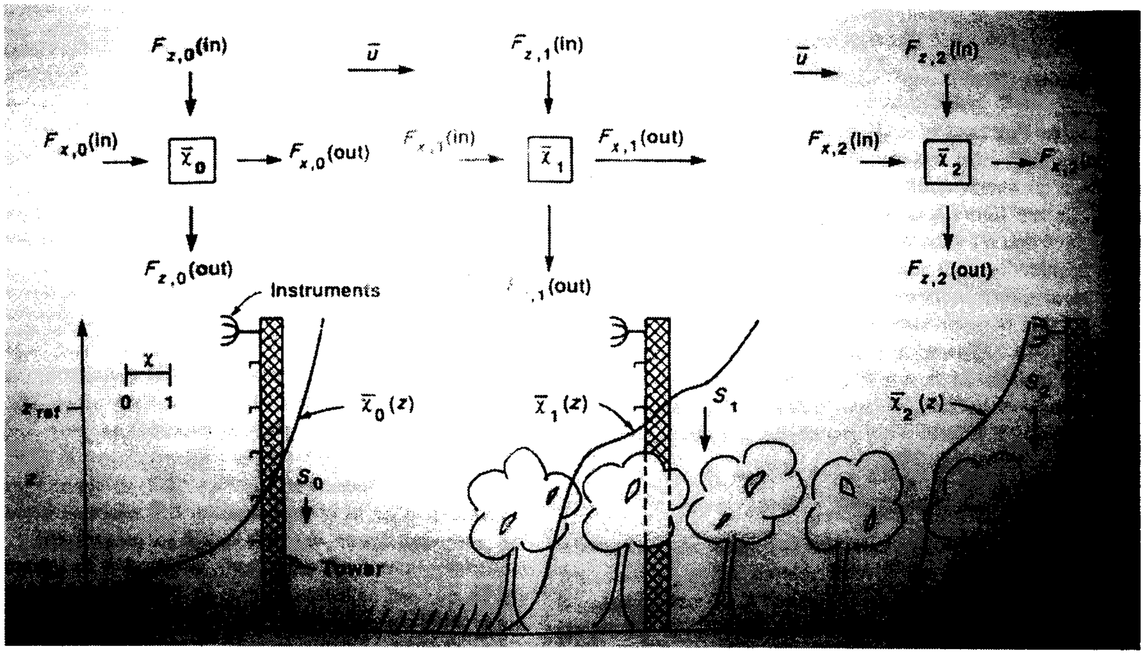
▷ **Theoretical background**

Direct flux measurement method: Measuring the flux of a scalar or momentum at a point centered on instruments placed at some height above the surface

Flux identification: The instruments must be located in the internal boundary layer where the flux is constant with height

Conservation equation for a scalar quantity in a control volume placed over the surface:

$$\frac{\partial \bar{s}}{\partial t} + \frac{\partial \overline{u_i s}}{\partial x_i} - D\frac{\partial^2 \bar{s}}{\partial x_i^2} = S$$



(the time rate of change of the mixing ratio at a fixed point in space + flux divergence or convergence - molecular diffusion = source/sink strength)

A constant flux layer requires stationarity:

$$\partial \bar{s}/\partial t = 0$$

The absence of sources or sinks between the surface and the instruments:

$$S = 0$$

Over a horizontally homogeneous surface for a sufficiently long upwind area, the mean vertical velocity or the dry air flux will be zero (1: longitudinal, 2: lateral):

$$\partial\left(\overline{u_i s}\right)/\partial x_i = D\left[\left(\partial^2 \bar{s}\right)/\left(\partial x_i^2\right)\right] = 0 (i = 1, 2)$$

$$\frac{\partial \overline{ws}}{\partial z} - D\frac{\partial^2 \bar{s}}{\partial z^2} = 0$$

Integrating above equation and summing the Reynolds convention yields

$$F_0 = D\left(\frac{\partial \bar{s}}{\partial z}\right)_0 = \left(\overline{w's'}\right)_z = F_z$$ (molecular diffusion at the underlying soil and leaf surface = turbulent eddy flux at height z)

The mean vertical turbulent flux of material over a horizontally homogeneous surface under steady-state conditions

$$F = -\overline{\rho_a w'c'}$$

This flux is directed downward when $F < 0$ and is directed upward when F is positive.

Under non-steady state and horizontally heterogeneous conditions

(flux = canopy storage + eddy flux + mass flow component arising from horizontal flow convergence/divergence or a non-zero mean vertical velocity at z_r)

$$F = \int_0^{z_r} \frac{\partial \bar{c}}{\partial t} dz + \left(\overline{w'c'}\right)_r + \bar{w}_r'\left(\bar{c}_r - \frac{1}{z_r}\int_0^{z_r} \bar{c} dz\right)$$

First equation of right hand, During daytime periods the CO_2 mixing ratio rarely varies. however larger changes in CO_2 may occur at dusk and dawn and make flux measurements during these periods unreliable.

Eddy covariance measurements made over non-ideal sites have value, too, even though annual estimates of net CO2 exchange may be error prone. Flux measurements from complex sites can provide information on the relationship between carbon fluxes and phenology, they can quantify how stand-scale carbon fluxes respond to environmental perturbations and they can quantify the factors causing year-to-year variability in carbon fluxes

▷ **Measurement criteria**

<u>Fetch to height ratio</u> (Gash, 1986):

advective effects are manifested when the internal boundary layer has not fully adjusted. The effective fetch for a 90% adjustment of the internal boundary layer in neutral stability can be computed as :

$$x_i = \frac{z\left(\ln\frac{z}{z_0} - 1 + \frac{z_0}{z}\right)}{k^2 \ln(0.90)}$$

<u>Complex terrain and sloping hills</u> (Finnigan 1983):

Because of negating the validity of the one-dimensional framework used to measure turbulent fluxes, a more complex experimental design than the one-dimensional framework discussed above is required to make flux measurements in complex terrain and sloping hills situations.

Turbulent wake effect above the tall or rough canopies:
general vegetation (Kaimal and Finnigan, 1994)
out of roughness sublayer
closed forest canopy (Raupach et al., 1980)
$h + 1.5L_t$
scattered open forest canopy (Garratt, 1978)
$3h$ or $4h$

Sensor response time:

$$f = \frac{n(z-d)}{u}$$

ability to respond to the full range of flux-carrying eddies is $0.001 < f < 10$, $n = u / l$ (reported by McBean, 1972; Anderson et al., 1986).

Sensor separation (Kristensen and Fitzjarrald, 1984):
The desired separation distance between two sensors should be less than the length scale of the smallest eddy to be detected, because the signals of two sensors become increasingly uncorrected with increasing separation distance.

$$d_1 \leq \frac{z-d}{5}$$

Sampling time (Lumley and Panofsky, 1964; Auble and Meyers 1992):

$$T_s = \frac{2\tau\sigma_t^2}{(\alpha\bar{\xi})^2} \approx 1000\frac{z-d}{u}$$

Flux covariance:
Assessment of the flux covariance requires that we sample the cospectrum of turbulent motions that exist in the atmosphere

$$\overline{w'c'} = \int_0^\infty S_{wc}(\varpi)d\varpi$$

$$\overline{w'c'}_{measured} = \int_0^\infty H(\varpi)S_{wc}(\varpi)d\varpi$$

High pass filtering: slow response time of sensor, long sensor path, slow sampling rate, or sensor separation

Low pass filtering: averaging method or sampling duration

▷ **Measurement correction**

<u>Density fluctuations error</u> (Webb et al., 1980):

Over uniform surfaces with active exchanges of heat and water vapor, this assumption is not always true. The exchanges of these entities lead to fluctuations in the density in dry air, which introduces a small but significantly non-zero mean vertical velocity.

$$F_c = \overline{w'\rho_c'} + \frac{m_a}{m_v}\frac{\bar{\rho}_c}{\bar{\rho}_a}\overline{w'\rho_v'} + \left(1 + \frac{\bar{\rho}_v}{\bar{\rho}_a}\frac{m_a}{m_v}\right)\frac{\bar{\rho}_c}{\bar{T}}\overline{w'T'}$$

<u>Tilt error</u> (Hyson et al., 1977; Wesely, 1970):

It is ofter impossible to orient the vertical velocity sensor so that the mean velocity is nearly zero, or find a perfectly flat experimental site. The influence of sensor tilt or terrain irregularities can contaminate the computation of the flux convariance by causing an apparent mean vertical velocity.

$$\overline{w'c'} = \overline{w'c_i'}\cos\theta - \overline{u'c_i'}\sin\theta\cos\sum \; - \overline{v'c_i'}\sin\theta\sin\sum$$

$$\cos\theta = \frac{\sqrt{\bar{u}^2 + \bar{v}^2}}{\sqrt{\bar{u}^2 + \bar{v}^2 + \bar{w}^2}}$$

$$\sin\theta = \frac{\bar{w}}{\sqrt{\bar{u}^2 + \bar{v}^2 + \bar{w}^2}}$$

$$\cos\sum \; = \frac{\bar{u}}{\sqrt{\bar{u}^2 + \bar{v}^2}}$$

$$\sin\sum \; = \frac{\bar{v}}{\sqrt{\bar{u}^2 + \bar{v}^2}}$$

▷ **Quality assessment** (Foken and Wichura, 1996)

<u>Instationarity test</u>:

$$\overline{x_i'x_j'} = \frac{1}{N/M}\left[\sum_{I=1}^{N/M}\overline{x_{iI}'x_{jI}'}\right]$$

$$\overline{x_i'x_j'} = \frac{1}{N-1}\left[\sum_{I=1}^{N/M}\sum_{k=1}^{M}x_{ik}\cdot x_{jk} - \frac{1}{N}\left(\sum_{I=1}^{N/M}\sum_{k=1}^{M}x_{ik}\right)\cdot\left(\sum_{I=1}^{N/M}\sum_{k=1}^{M}x_{jk}\right)\right]$$

If there is a difference of less than 30% between the covariances or dispersions determined with above two equations, then the measurement is considered to be stationary.

Correlation coefficient:

Typical correlation coefficients for momentum and sensible heat flux

Reference	Friction velocity	Sensible heat flux
Hicks (1981)	-0.32	+0.35 for z/L -> -0.0
		+0.6 for z/L -> -2.0
Kaimal et al. (1990)	-0.3	+0.5 for $z/L < 0.0$
Kaimal and Finnigan (1994)	-0.35	+0.5 for $-2 < z/L < 0$
		-0.4 for $0 < z/L < 1$

Integral turbulence characteristics:

$$\frac{\sigma_w}{u_*} = a_1 \cdot \left[\varphi_m(z/L)\right]^{b_1}$$

$$\frac{\sigma_T}{T_*} = a_2 \cdot \left[(z/L) \cdot \varphi_h(z/L)\right]^{b_2}$$

Dependence of the integral turbulence characteristics on the stratification according to Foken et al. (1991)

z/L	*sigma* w/U_*	*sigma* u/U_*	*sigma* T/T_*
< -1	$2.00(-z/L)^{1/6}$	$2.83(-z/L)^{1/6}$	$1.00(-z/L)^{-1/3}$
-1 ... -0.0625	$2.00(-z/L)^{1/8}$	$2.83(-z/L)^{1/8}$	$1.00(-z/L)^{-1/4}$
0.0625 ... < 0	1.41	1.99	$0.50(-z/L)^{-1/2}$

▷ **Systematic bias errors** (Baldocchi, 2003)

Lack of energy balance closure:

Filtering of low frequency flux contributions

Advection

Different footprints viewed by the eddy flux and the available energy measurement systems

Independent test: lysimeter or watershed water balances

Nighttime bias errors when winds are light and intermittent tend to produce an underestimate in the measurement of ecosystem respiration

Temperature dependent respiration function

Measured during windy periods using a regression between CO_2 flux density and friction velocity

Comparative analysis

Model, chamber and soil and biomass inventories

Biogeochemical models and remote sensing indices

Partition net carbon fluxes into the components

▷ **Data gap filling** (Falge et al., 2001)

Mean diurnal variation

Semi-empirical methods

- Look-up tables
- Nonlinear regression methods

statistical and empirical model

▷ **Flux footprint** (Schuepp et al., 1990)

The relative contribution to the vertical flux at height *z*, coming from an infinite crosswind source of unit width at an upwind distance x, in neutral conditions is given by:

$$\frac{1}{Q_0}\frac{dQ}{dx} = \frac{u(z-d)}{u^* k x^2} e^{-u(z-d)/(k u^* x)}$$

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