Chapter 4
Corrections and Data Quality Control

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This chapter describes corrections that must be applied to measurements because practical instrumentation cannot fully meet the requirements of the underlying micrometeorological theory. Typically, measurements are made in a finite sampling volume rather than at a single point, and the maximum frequency response of the sensors is less than the highest frequencies of the turbulent eddies responsible for the heat and mass transport. Both of these cause a loss of the high-frequency component of the covariances used to calculate fluxes. Errors also arise in calculating fluxes of trace gas quantities using open-path analyzers because of spurious density fluctuations arising from the fluxes of heat and water vapor. This chapter gives the reader an overview of how these sources of error can be eliminated or reduced using some model assumptions and additional measurements. Corrections needed for some specific instruments are presented (Sect. 4.1), followed by a discussion...
of the generally observed lack of closure of the energy balance using the sum of latent and sensible heat fluxes (Sect. 4.2). The chapter closes with a discussion of measures needed to determine the quality of the final calculated fluxes (Sect. 4.3).

4.1 Flux Data Correction

4.1.1 Corrections Already Included into the Raw Data Analysis (Chap. 3)

In this chapter, we assume that several steps of preprocessing of the raw velocity and scalar time series have been completed (Sect. 3.2.2). These include despiking of the raw data (Højstrup 1993; Vickers and Mahrt 1997) and cross-correlation of the time series data in order to shift all signals to the same time base. The most important issues are the delay time of closed-path sensors (Sect. 3.2.3.2) and digitization delays in the sensors. The acoustic temperature measured by sonic anemometers must be corrected for crosswind influences (Schotanus et al. 1983; Liu et al. 2001). Current sonic anemometers include this correction in their firmware, but this is not the case for some older anemometers (Sect. 2.3.2) It is also assumed that the coordinate system has been rotated to ensure zero vertical wind velocity over a certain averaging time (Sect. 3.2.4). This rotation is called a tilt correction (Tanner and Thurtell 1969; Hyson et al. 1977) or coordinate rotation (Kaimal and Finnigan 1994) and can be applied to two or three axes, where the three-axis rotation is not recommended (Finnigan et al. 2003). Currently the planar-fit method (Wilczak et al. 2001) is often used, which overcomes some deficiencies of the double rotation for short averaging times. For more details see Chap. 3.

4.1.2 Conversion of Buoyancy Flux to Sensible Heat Flux (SND-correction)

Conversion of buoyancy flux to sensible heat flux is known as SND-correction after the three authors of the paper Schotanus et al. (1983), formerly also called the Schotanus-correction. It is based on the transformation of sonic or acoustic temperature \( \theta_s \) to actual air temperature using (see also Chap. 3, Eq. 3.3)

\[
\theta_s = \theta \left( 1 + 0.32 \frac{e}{p} \right)
\]  

(4.1)

where \( p \) is atmospheric pressure and \( e \) is the partial pressure of water vapor. Extra measurements of humidity are thus needed to calculate the sensible heat flux \( H = \rho c_p \theta' \) while \( \rho c_p \theta'_s \) is the buoyancy flux (Kaimal and Gaynor 1991).
Application of Reynolds decomposition to Eq. 4.1 and computation of the temperature variance and covariances lead to the following relations (Schotanus et al. 1983)

$$\sigma_{\tilde{\theta}}^2 = \sigma_{\tilde{\theta}_s}^2 - 1.02 \tilde{\theta} \chi' \tilde{\theta}' - 0.51^2 \chi' \tilde{\theta}^2$$  \hspace{1cm} (4.2)

$$\overline{w' \tilde{\theta}'} = \overline{w' \tilde{\theta}'_s} - 0.51 \tilde{\theta} \overline{w' \chi'_{\nu}},$$  \hspace{1cm} (4.3)

where the factor 0.51 results from the multiplication of factor 0.32 in Eq. 4.1 by the ratio of dry air and water molar masses. This correction is straightforward if the latent heat flux is also measured. If such measurements are not available, an estimate of the Bowen ratio $Bo = H / \lambda E$, for example, from the temperature and moisture difference between two levels, can be used (Arya 2001; Hatfield and Baker 2005; Foken 2008b; Monteith and Unsworth 2008). In this case, the sensible heat flux is calculated using ($\lambda$: heat of evaporation):

$$H = \rho c_p \overline{(w' \tilde{\theta}')} = \rho c_p \frac{(\overline{w' \tilde{\theta}'_s})}{1 + \frac{0.51 \epsilon_p \theta}{\lambda Bo}} $$  \hspace{1cm} (4.4)

Because the Bowen-ratio involves the sensible and latent heat flux, both must be known to apply this method. Thus, the solution of this equation often is made iteratively to update the sensible heat flux in the Bowen-ratio. However, Oncley et al. (2007) show that the equations can be solved simultaneously using the two Eqs. 4.3 and 4.4 and the calculation of both corrected fluxes of sensible and latent heat.

Note that the buoyancy flux required, for example, in the computation of the Obukhov length (Foken, 2006). The buoyancy flux is $w' \tilde{\theta}'_{\nu}$ which is close to, but not exactly the same as, $\overline{w' \tilde{\theta}'_s}$ (see Sect. 3.2.1.1). When $Bo$ is large, $\overline{w' \tilde{\theta}'_s} \approx \overline{w' \tilde{\theta}'}$ and no correction to $\overline{w' \tilde{\theta}'}$ is needed. However, in the general case, it is necessary to use Eq. 4.3 and then compute $\overline{w' \tilde{\theta}'_{\nu}}$ from $\overline{w' \tilde{\theta}'} + 0.51 \tilde{\theta} \overline{w' \chi'_{\nu}}$.

### 4.1.3 Spectral Corrections

#### 4.1.3.1 Introduction

Eddy covariance systems, like all measuring instruments, act as filters, removing both high- and low-frequency components of the signal. High-frequency losses are mainly due to inadequate sensor frequency response, line averaging, sensor separation and, in closed-path systems, air transport through the tubes. The impact of high-frequency losses on cospectral density are illustrated schematically in Fig. 4.1 (see also Sect. 1.5). Correction procedures for high-frequency losses are
Fig. 4.1 Normalized turbulence spectra for an ideal instrument, which measures the unaffected turbulence spectra, and a nonideal instrument. The missing energy between both response curves must be corrected (n normalized frequency; f frequency, \( z \) height, \( u \) wind velocity, \( S_{xx} \): energy density of the parameter \( x \); \( \sigma_x^2 \): dispersion of the parameter \( x \)).

Described in Sect. 4.1.3.2. Low-frequency losses result from the finite sampling duration, with the averaging period not always being sufficiently long to include all relevant low frequencies. The use of detrending or recursive filtering, by attenuating fluctuations at periods larger than the filter time constant, may enhance this effect. They are therefore not recommended in general. Their use could however be necessary, when a sensor calibration drift should be removed. In this case, all information relative to fluctuations at period larger than recursive filter time constant would be lost.

### 4.1.3.2 High-Frequency Loss Corrections

The relative error on the flux introduced by high-cut (often referred to as low-pass) filtering was described in Chap. 1, Eq. 1.31 by the formula:

\[
\delta_s / F_s^{EC} = 1 - \frac{\int_0^\infty \frac{C_{ws}(f)}{T_{ws}(f)} df}{\int_0^\infty \frac{C_{ws}(f)}{df}}.
\]

(4.5)

where \( C_{ws} \) represents the nonfiltered or ‘ideal’ cospectral density and \( T_{ws} \) represents the transfer function of the system, involving high-cut and, possibly, low-cut (high-pass) filtering effects. This section is focused on the computation of the correction.
for high-cut filtering effects. This could be evaluated if both the system transfer function $T_{ws}(f)$ and the nonfiltered cospectrum $C_{ws}(f)$ are known.

The effects of high-frequency losses on spectral and cospectral density are illustrated in Fig. 4.2. Poor high-frequency response of the (closed-path) measurement system causes the CO$_2$ cospectrum ($C_{wc}$) to roll-off more quickly than for temperature ($C_{w\theta}$), and the roll-off is even more rapid for the water vapor cospectrum ($C_{wv}$) due to its adsorption/desorption on tubing walls. Reduced high-frequency response causes the normalized $C_{wc}$ cospectrum to be lower than that for the normalized $C_{w\theta}$ cospectrum, and the effect is even greater for the $C_{wv}$ cospectrum. This causes the fluxes of CO$_2$ and water vapor to be underestimated.

The following section describes two approaches commonly used to correct for imperfect instrument high-frequency responses.

In the theoretical approach, the transfer function $T_{ws}$ is deduced from prior knowledge of the measuring system and the cospectral function $C_{ws}$. In the experimental approach, the fractional error is computed using the normalized ratio of two cospectral density functions measured simultaneously at the same site: one
referring to the filtered scalar of interest, the other considered as a reference and supposedly real cospectrum. In practice, the sensible heat cospectrum is usually used for the second. A description of these approaches, their implementation, their respective advantages and disadvantages are discussed below. Many applicants use the more general theoretical approach while the experimental approaches need site- and sensor-specific investigations.

The Theoretical Approach

This approach was first devised for eddy covariance systems by Moore (1986) and later extended to CO\textsubscript{2} closed-path systems, in particular by Leuning and Moncrieff (1990), Leuning and King (1992), Lee and Black (1994), Leuning and Judd (1996), Massman (2000), Ibrom et al. (2007\textsubscript{a}), Massman and Ibrom (2008), and Horst and Lenschow (2009). An addition considering the phase shift of a low-pass filter was proposed by Horst (2000) and Massman and Ibrom (2008).

The total transfer function \( T_{ws} \) describing an eddy covariance system for the vertical flux of a tracer \( s \) may be described as a function of real frequency \( f \):

\[
T_{ws}(f) = G_w(f) \cdot G_x(f) \cdot T_{ss}(f) \cdot \sqrt{T_{pw}(f)} \cdot \sqrt{T_{ps}(f)} \cdot \sqrt{T_{ta}(f)}
\]  

(4.6)

where:

\( G_w(s)(f) \) describe the high frequency loss by the sensors and are defined as

\[
G_w(s) = \left[ 1 + \left( 2\pi f \tau_{w(s)} \right)^2 \right]^{-1/2}
\]  

(4.7)

in which \( \tau_{w(s)} \) are time constants, specific to the sensors (Moore 1986), see also Horst (1997)

\( T_{ss}(f) \) describes the high-cut filtering due to lateral separation between two instruments (in this case the sonic anemometer and the open-path analyzer or the inlet of the closed-path analyzer). Moore (1986) gives a simple expression for this displacement, based on empirical fits to an isotropic turbulence model:

\[
T_{ss}(n) = e^{-9.9n_{ss}^{1.5}}
\]  

(4.8)

where:

\[
n_{ss} = \frac{f d_{ss}}{\overline{u}}
\]  

(4.9)

Here, \( \overline{u} \) is the average wind speed and \( d_{ss} = d_{sa} |\sin(\beta_d)| \) is the effective lateral separation distance between the two sensors. \( d_{sa} \) is the actual separation distance and \( \beta_d \) is the angle between the line joining the sensors and the wind direction. However, only nonisotropic turbulence can produce a flux. For this reason, Horst
and Lenschow (2009) created a more realistic nonisotropic turbulence model that depends on stability from a unique set of field observations and used it to obtain somewhat more involved expressions for $T_{ss} (n_s)$. 

- $T_{pw}$ represents the transfer function for the wind vector component line averaging. It can be approximated by (Kaimal et al. 1968; Horst 1973):

$$T_{pw}(n) = \frac{2}{\pi n_w} \left( 1 + \frac{e^{-2\pi n_w}}{2} - 3 \frac{1 - e^{-2\pi n_w}}{4\pi n_w} \right)$$  \hspace{1cm} (4.10)

where

$$n_w = \frac{f d_{pl}}{u}$$  \hspace{1cm} (4.11)

and $d_{pl}$ represents the sonic anemometer path length.

- $T_{ps}$ represents the transfer function for the scalar line averaging (Moore 1986):

$$T_{ps}(n) = \frac{1}{2\pi n_s} \left( 3 + e^{-2\pi n_s} - 4 \frac{1 - e^{-2\pi n_s}}{2\pi n_s} \right)$$  \hspace{1cm} (4.12)

where:

$$n_s = \frac{f d_s}{u_{pl}}$$  \hspace{1cm} (4.13)

$u_{pl}$ is the average air speed inside the sensor path length and $d_s$ represents the infrared gas analyzer (IRGA) path length or the path length of the sonic anemometer for temperature measurements.

- $T_{ta}$ describes the fluctuation attenuation due to air transport in the tubes of closed path systems. When flow in the tube is laminar, Lenschow and Raupach (1991) and Leuning and King (1992) proposed the following formulation for passive scalars such as CO$_2$:

$$T_{ta} = \exp \left\{ -\frac{\pi^3 r_t^4 f^2 L_t}{6 D_s Q} \right\}$$  \hspace{1cm} (4.14)

or when the flow in the tube is turbulent:

$$T_{ta} = \exp \left\{ -160 \cdot Re^{-1/8} \cdot \frac{\pi^2 r_t^5 f^2 L_t}{Q^2} \right\}$$  \hspace{1cm} (4.15)

Here, $r_t$ and $L_t$ are the tube radius and length, $Q$ is the volumetric flow rate in the tube, and $D_s$ is the molecular diffusivity of scalar $s$. The Reynolds number is defined as $Re = \frac{2Q}{\pi r_t v}$, in which $v$ is the kinematic viscosity. Turbulent flow in tubing occurs when $Re \gtrsim 2300$. 

Massman and Ibrom (2008) recently reexamined these formulations and found that Eq. 4.15 tends to underestimate the attenuation of fluctuations. For passive scalars, they proposed the alternative expression

$$T_{ta} = \exp \left\{ -\left(160 \cdot Re^{-1/8} + 2666 \cdot Re^{-29/40} \right) \frac{\pi^2 r_l^5 f^2 L_t}{Q^2} \right\}$$  \hspace{1cm} (4.16)$$

while for scalars adsorbed/desorbed on the walls of the tubing they proposed

$$T_{ta} = \exp \left\{ -\left(160 \cdot Re^{-1/8} + 2666 \cdot Re^{-29/40} + 8000 \cdot Sc^{-1/2} \right) \times \left[10^9 \cdot Re^{-2 \cdot r_h \cdot e^{l^* r_h}} \right] \right\} \frac{\pi^2 r_l^5 f^2 L_t}{Q^2}$$  \hspace{1cm} (4.17)$$

In (4.17), $Sc = v / D_s$ is the Schmidt number, $r_h$ the relative humidity, and $l^*$ an empirically determined coefficient set to 8.26 by Massman and Ibrom (2008).

Some additional remarks about the use of the preceding equation set:

- All transfer functions must be expressed in terms of the real frequency $f$ before being used with Eq. 4.6.
- Equation 4.8 can only be used in the unstable case, and if the sensor separation is less than 10% of the aerodynamic measuring height (height above zero-plane displacement). Under stable stratification, the distance between the sensors should not be greater than 0.7% of the Obukhov length (Moore 1986).
- The relation describing line averaging of vector quantities is more complex because of the influence of the sensor geometry (Kaimal et al. 1968; Horst 1973). Equation 4.10 is an approximation that is accurate to about 2% for the vertical wind component.
- Equation 4.12 is an approximation when the angle between the line-averaging path of the sensor and the wind field is 90° (Gurvitch 1962). For all angles, the equation is given by Silverman (1968).
- A correction of the longitudinal sensor separation is only necessary if the covariance was not maximized by cross correlation analysis (see Sect. 3.2.3.2, Mauder and Foken 2004). The transfer function for lateral separation can also be used for the correction of longitudinal separation, (Moore 1986) since in both cases the 3 dB-point (damping of the signal by $1/\sqrt{2}$) is the same in both transfer functions.
- Implementation of the theoretical approach needs a spectral model. For unstable conditions, we recommend using spectra presented by Højstrup (1981) for wind components, those of Kaimal et al. (1972) for scalar spectra, and those of Kristensen et al. (1997) for scalar cospectra. Other models for spectra and cospectra are given in textbooks (Kaimal and Finnigan 1994; Foken 2008b).
- A careful application of Eqs. 4.5–4.17 shows that for open-path sensors, the high-frequency spectral correction is not directly dependent on windspeed; $\delta_s / F_{EC}^s$ depends only on the ratios of the various length scales, $d_{ss}$, $d_{pl}$ etc. divided by
The correction factor is weakly dependent on stability under unstable and neutral conditions because there is then no significant variation in the shape of the normalized cospectrum (Kaimal et al. 1972). However, the normalized cospectrum does vary significantly under stable conditions and this causes $\delta_s/F_s^EC$ to vary strongly with stability.

There is a wind speed dependence of $\delta_s/F_s^EC$ for closed-path systems when the flow in the tube is laminar. In this case, the half-power frequency of filtering by the tubing $f_{0,t}$, is less than that for the high-frequency roll-off of atmospheric turbulence: $f_{0,s} = (2 \sim 5) \bar{u}/(h_{in} - d)$. The attenuation increases with wind speed in this case. If flow in the pipe is sufficiently turbulent, $f_{0,s}$ will be greater than $f_c$ and there will be no dependence of the attenuation on wind speed. The transfer function for flow of passive scalars through tubing is a function of volumetric flow rate, tube length, and radius, while adsorption isotherms need to be considered for sorbing scalars.

The theoretical approach has the advantage of relying on a fundamental description of the system and allows a comprehensive description of filtering processes. However, it also has some shortcomings. First, the theoretical cospectral densities that are proposed may not correspond to the real cospectral densities observed at the sites. Amiro (1990) showed notably that at forested sites, the measured sensible heat cospectral density diverged from the Kaimal et al. (Kaimal et al. 1972) cospectra in the inertial range. This was confirmed experimentally by de Ligne et al. (2010). It is also well known that the low-frequency part of the cospectrum is not universally defined in unstable conditions and could depend on mesoscale movements specific to a given site. Secondly, all processes cannot be thoroughly described by the transfer function, especially for closed-path systems. Indeed, in this case, problems may come from uncertainties in volume flow, if the systems do not employ mass-flow controllers; from uncertainties in the flow regime, the Reynolds number value being not always a sufficient criterion for turbulent flow in the tubing; or from the impact of particle filters on the transfer function (Aubinet et al. 2000; Aubinet et al. 2001). In addition, the transfer functions for tube attenuation (4.14–4.15) are exact only in the case of straight horizontal tubes, which is rarely the case in eddy covariance systems.

In practice, if the theoretical approach is well suited to open-path systems, its application to closed-path systems appears more problematic. In any case, it is recommended to compare the theoretical transfer functions obtained with the set of Eqs. 4.5–4.17 with experimental transfer functions obtained with the procedure described in the next section.

The Experimental Approach

The experimental approach assumes that we can measure without significant error the cospectrum of one quantity (e.g. $\overline{w^2\theta^2}$) and that this can be used to rescale the cospectrum of another quantity that is subject to high-cut filtering. The computation process consists of: (1) Selecting long time periods (3 h at least in order to reduce the uncertainties on the low-frequency part of the cospectra) with sunny, stationary
conditions, with different wind velocities (and different air humidity, as far as sorbing scalars are concerned); (2) for each period, calculating the cospectra for heat and for the scalar of interest and, finally, (3) calculating the transfer function as the ratio of the normalized cospectral densities as

\[ T_{ws}^{\exp}(f) = \frac{N_\theta C_{ws}(f)}{N_s C_{w^\theta}(f)} \]  \hspace{1cm} (4.18)

where \( N_\theta \) and \( N_s \) are normalization factors. Similarity would require that \( \frac{N_\theta}{N_s} = \frac{\omega_\theta/\theta}{\omega_s/\theta_s} \) but, as covariances are affected by high-frequency attenuation, they cannot be calculated exactly. Aubinet et al. (2000) proposed thus to compute them as

\[ \frac{N_\theta}{N_s} = \frac{\int_{0}^{f'} C_{w^\theta}^{\exp}(f) df}{\int_{0}^{f'} C_{ws}^{\exp}(f) df} \]  \hspace{1cm} (4.19)

where the limit frequency, \( f' \), is high enough to allow computation of the normalization factor with enough precision, and low enough, to not be affected by high-frequency attenuation (Aubinet et al. 2000).

For each period, a sigmoidal function may be fitted to \( T_{ws}^{\exp}(f) \), from which a half-power frequency \( f_{o,s} \) in Hz may be computed. The half-power frequency (or its relation with wind velocity and, if necessary, air humidity in the case of closed paths with laminar flows) can then be deduced and further used to compute the transfer function response to wind velocity.

Different sigmoidal functions are used to fit the experimental transfer functions. The most commonly used equations are the Gaussian (Aubinet et al. 2001):

\[ T_{ws}^{fit}(f) = \exp \left[ -\ln(2) \left( \frac{f}{f_{o,s}} \right)^2 \right] \]  \hspace{1cm} (4.20)

and the Lorentzian (Eugster and Senn 1995):

\[ T_{ws}^{fit}(f) = \frac{1}{1 + \left( \frac{f}{f_{o,s}} \right)^2} \]  \hspace{1cm} (4.21)

Eq. (4.21) can also be considered as characteristic of a first-order filter performed by a resistor-inductor (RL) circuit of unitary resistance and inductance equal to \( L_{self} \) in Hz\(^{-1}\). Under these conditions, the filter inductance is related to the half-power frequency by:

\[ f_{o,s} = \frac{1}{2\pi L_{self}} \]  \hspace{1cm} (4.22)
This equation is widely used, for example, by Horst (1997), Su et al. (2004), Ibrom et al. (2007a), Hiller et al. (2008), and Mammarella et al. (2009). However, de Ligne et al. (2010) found that the shape of Eqs. 4.20 and 4.21 was not well suited to water vapor transfer functions, their decrease with frequency being too sharp. They proposed an alternative relation:

\[
T_{ws}^{fit}(f) = \exp \left[ -\ln(2) \left( \frac{f}{f_{o,s}} \right)^n \right]
\] (4.23)

where the parameter \( n \) was lower than two and varied (like \( f_{o,s} \)) with air saturation deficit. Other procedures suited to water vapor transfer functions were proposed by Ibrom et al. (2007a) and Mammarella et al. (2009).

Similar to the theoretical method, implementation of the experimental approach also needs a spectral model. Højstrup (1981), Kaimal et al. (1972) or Kristensen et al. (1997) models could be used as above. Alternatively, the use of experimental, site-specific, cospectral models could be relevant above forests (de Ligne et al. 2010). The introduction of the experimental transfer function and of the cospectral model (e.g., Eq. 1.30b) in Eq. 4.5 will lead to a correction function which has the same properties as those obtained with the theoretical approach, that is, it is a single function of the wind speed and measurement height in unstable conditions and of the wind speed, measurement height, and stability in stable conditions. It is stable as long as the set up remains unchanged and could be applied to every individual flux estimate.

The experimental approach relies on different hypotheses: First, it assumes that the processes of atmospheric turbulent transport of sensible heat and other tracers are similar and therefore the cospectral densities should be proportional to each other. This hypothesis was proposed notably by Wyngaard and Coté (1971), Panofsky and Dutton (1984), and Othaki (1985), and it has been widely used in spectral correction schemes. It was tested above forests by various authors (Anderson et al. 1986; Monji et al. 1994; Ruppert et al. 2006) who found high scalar similarity at the midday period.

Secondly, it assumes that the high-frequency attenuation of sensible heat cospectral density is negligible compared with those affecting other tracers (i.e., that fluctuation attenuation due to electronic response time or path averaging takes place at much higher frequencies than attenuation due to mixing in the tube and from sorption/desorption). If this hypothesis clearly makes sense in closed-path systems where the high-frequency losses are mainly due to tube attenuation and sensor separation, it is less relevant in the case of open-path systems. As a result, the experimental approach could lead to an underestimation of the high-frequency correction in open-path systems and would not be recommended in this case (Aubinet et al. 2000; Aubinet et al. 2001).
4.1.3.3 Low-Cut Frequency

The low-cut frequency correction is necessary if very large eddies occur that are not completely sampled over the averaging period. Most researchers use an averaging period of 30 min to calculate eddy fluxes, but this may not be long enough to capture all of the low-frequency contributions to the fluxes (Finnigan et al. 2003). It is therefore best to test if the flux has its maximum value within the adopted averaging time. This is done using the so-called ogive test (Desjardins et al. 1989; Oncley et al. 1990; Foken et al. 1995). The ogive ($Og_{ws}$) is calculated using the cumulative integral of the cospectrum of the turbulent flux beginning with the highest frequencies (Fig. 4.3):

$$Og_{ws}(f_0) = \int_{f_0}^{\infty} C_{ws}(f) \, df$$

The averaging period is satisfactory if the value of the integral approaches a constant value (the flux) at low frequencies.

Fig. 4.3 Converging ogive ($Og_{ws}$) and cospectrum ($f \cdot CO_{ws}$) of the sensible heat flux during the LITFASS-2003 experiment (17.06.2003, 12:30–16:30 UTC, Foken et al. 2006)
Foken et al. (2006) have shown for the LITFASS-2003 experiment (Mengelkamp et al. 2006) that in about 80% of all cases, the ogive converged within a period of 30 min. In the remaining cases, mainly in the transition periods of the day, the ogives did not converge or reached a maximum value before the integration time of 30 min, and then decreased in magnitude. In these cases, it would be best to recalculate the fluxes using a different averaging time (longer for the first case and shorter for the second). Since this change of averaging time is difficult to implement in operational data streams, the ogive method is typically used on data from research campaigns. For practical application it may be helpful to check selected time series for unstable, stable, and transition periods and to apply the findings for the data set.

4.1.4 WPL Corrections

4.1.4.1 Introduction

This correction was formerly called the Webb-correction after the first author of a conference paper to correct water vapor fluxes, but now is called WPL-correction after the three authors (Webb, Pearman and Leuning 1980, WPL) who extended its use to measurement of eddy fluxes of trace gases. After the first publication by Webb et al. (1980), several authors discussed this problem, summarized by Fuehrer and Friehe (2002), and Lee and Massman (2011) with different approaches (Bernhardt and Piazena 1988; see Liebenthal and Foken 2003; Liebenthal and Foken 2004) who found equal results but also controversial solutions (Liu 2005). A clarification of the problem was recently given by Leuning (2004, 2007). The correction is necessary because fluctuations in temperature and humidity cause fluctuations in trace gas concentrations that are not associated with the flux of the trace gas we wish to measure. The correction to the measured flux can be large, for example, the additive correction significantly reduces the CO₂ flux calculated using the covariance of vertical velocity and density (Fig. 4.4). A very careful application of this correction is essential for all trace gases as discussed below.

4.1.4.2 Open-Path Systems

Webb et al. (1980) derived the following expression for the eddy flux of a trace gas \( c \) to account for the effects of density fluctuations due to temperature and humidity fluctuations when measurements are made using open-path instruments:

\[
F_c(h_m) = \bar{w'} \bar{c'} + \mu \left( \frac{\bar{c'}}{\bar{d'}} \right) \bar{w'} \bar{c'} + (1 + \mu \sigma) \left( \frac{\bar{c'}}{\bar{\theta'}} \right) \bar{w'} \bar{\theta'}
\]  (4.25)
where $\mu = m_d/m_v$ is the ratio of molar masses of dry air and water vapor, and $\sigma = \bar{\rho}_v/\bar{\rho}_d$, the ratio of the densities of water vapor and dry air. The other quantities have been defined earlier.

To derive the last two terms on the right side of this fundamental equation, WPL assumed that there is no flux of dry air through a plane at the measurement height $h_m$. This assumption is correct for horizontally homogeneous, steady-state conditions, i.e. when there is no change in mass storage within the air layer below $h_m$. However, for non steady-state flows there is a net flux of all components of the air through the plane at $h_m$, thus violating the assumptions of the original paper and throwing doubt on the WPL equations.

This issue was resolved in Chap. 1 where it was shown that the one-dimensional conservation equation for the trace quantity $c$ under nonsteady-state conditions is given by Eq. 1.23:

\[
F_c = F_c(0) + \int_0^{h_m} S_c \, dz + \int_0^{h_m} \bar{\rho}_d \, w' \chi_c \, h_m + \int_0^{h_m} \bar{\rho}_d \, \frac{\partial \chi_c}{\partial t} \, dz + \int_0^{h_m} \left[ \chi_c(z) - \chi_c(h) \right] \frac{\partial \bar{\rho}_d}{\partial t} \, dz
\]

(4.26)

Our objective is to estimate the sum of the flux of $c$ at the ground (term I) plus the integrated contribution of all sources and sinks of $c$ below the measurement height $h_m$ (term II) by measuring the net turbulent flux at $h_m$ (term III), the change in
storage (term IV), and the mean flux of \( c \) caused by the change in density of dry air (term V). The eddy flux term III in Eq. 4.26 is identical to that derived by WPL for steady-state conditions and it is thus only necessary to add terms IV and V equation for nonsteady-state conditions.

Densities, rather than mixing ratios are typically measured using open-path eddy covariance instruments and following WPL, the eddy flux for a trace gas \( c \) is calculated using (Eq. 4.25).

The corresponding eddy flux of water vapor is

\[
\bar{F}_v(h_m) = (1 + \mu \sigma) \left[ \bar{w}' \rho_c' + \left( \frac{\bar{\rho}_v}{\bar{\theta}} \right) \bar{w}' \theta' \right]
\] (4.27)

### 4.1.4.3 WPL and Imperfect Instrumentation

The above theory assumes that the eddy fluxes are measured using perfect instrumentation. This requires the cospectral frequency response of the instrument array used to measure \( \bar{w}' \rho_c' \) and \( \bar{w}' \theta' \) to be identical to that used to measure \( \bar{w}' \rho_c' \). As discussed in Sect. 4.1.3, it is necessary to correct for any differences in frequency response of the instrument array before applying the WPL corrections. An extreme example is shown in Fig. 4.5 by measurements made by Kondo and Tsukamoto (2008) over an asphalt parking lot where the CO2 and water vapor fluxes were zero. In this case, the \( \bar{w}' \theta' \) correction term should be of equal magnitude but of opposite sign to the raw \( \bar{w}' \rho_c' \) in each frequency band. The error term arises because of a mismatch in the high-frequency components of the \( \bar{w}' \theta' \) and \( \bar{w}' \rho_c' \) covariances. The true CO2 flux is obtained by first adjusting the high-frequency component of the raw \( \bar{w}' \rho_c' \) covariance to the red line (Sect. 4.1.3) and then applying the WPL corrections.

### 4.1.4.4 Closed-Path Systems

There is often considerable loss of eddy flux data when open-path gas analyzers are used at sites where rain, mist, and snow impair measurements of trace gas concentrations. Closed-path gas analyzers provide an attractive alternative because of lower rates of data loss but such measurement systems require significantly different corrections for time delays, high-frequency filtering, and density effects compared to those needed for open-path systems (see Sects. 2.4, 4.1.4.2).

Temperature fluctuations in tubing with high thermal conductivity are reduced to 1% of their initial value when the tubing length to radius \( L_t/r_t > 1200 \) for laminar flow in the tube and when \( L_t/r_t > 500 \) for turbulent flow (Leuning and Judd 1996; Sect. 4.1.4.2). When temperature fluctuations at all frequencies are eliminated by the air sampling system there is no need for the \( \bar{w}' \theta' \) correction term in Eq. 4.27 and the fluxes for a trace gas and water vapor are then calculated using

\[
\bar{F}_c(h_m) = \left( \frac{\bar{\rho}_v}{\bar{\theta}} / \bar{\rho}_c \bar{\theta} \right) \left[ \bar{w}' \rho_{c,i}' + \mu \left( \frac{\bar{\rho}_{c,i}}{\bar{\rho}_{v,i}} \right) \bar{w}' \rho_{v,i}' \right]
\] (4.28)
Fig. 4.5 Cospectra for vertical velocity \( w \) and CO\(_2\) concentration \( c \), measured using a sonic anemometer and an open-path sensor located 0.25 m apart at a height of 1.62 m over an asphalt parking lot (adapted from Kondo and Tsukamoto 2008). Also shown are the WPL correction terms to the raw CO\(_2\) flux due to sensible and latent heat.

\[
F_v(h_m) = \left( \frac{\bar{p} \bar{\theta}_i}{\bar{p}_i \bar{\theta}} \right) \left[ (1 + \mu \frac{\bar{p}_{v,i}}{\bar{p}_{d,i}}) w' \rho'_{v,i} \right]
\]

(4.29)

where \( \bar{p} \) and \( \bar{\theta} \) are the mean pressure and absolute temperature in the ambient air and \( \bar{p}_i \) and \( \bar{\theta}_i \) are the corresponding quantities measured within the gas analyzer.

Complete elimination of temperature fluctuations may not be achievable for practical lengths of tubing with walls of low thermal conductivity, in which case some unknown fraction of the \( w' \theta' \) density correction must be applied. The solution is to measure the temperature and pressure fluctuations within the gas analyzer at the 10 or 20 Hz normally used for the rest of the eddy covariance system. Eddy fluxes of the trace gas and water vapor are then calculated using

\[
\overline{F}_c(h_m) = \bar{\rho}_d w' \chi'_c, \quad \overline{F}_v(h_m) = \bar{\rho}_d w' \chi'_v
\]

(4.30)
where the instantaneous mixing ratios $\chi'_c$ and $\chi'_v$ of $c$ and $v$ relative to dry air are given by

$$\chi'_c = \frac{\rho'_{c} / m_d}{p'_i / (R\theta'_i) - \rho'_{v} / m_v}, \quad \chi'_v = \frac{\rho'_{v} / m_d}{p'_i / (R\theta'_i) - \rho'_{v} / m_v}$$

(4.31)

Normalized CO$_2$ and water vapor cospectra shown in Fig. 4.2, d were calculated using these formulas. We note that some closed-path instruments measure the wall temperature of the gas analysis cell rather than the desired air temperature and use of these measurements thus introduces some high-cut filtering into the $\theta'$signal.

The rapid attenuation of fluctuations in CO$_2$ and water vapor at high frequencies seen in Fig. 4.2b causes a corresponding loss of high-frequency covariance. The resultant underestimate of the fluxes is proportional to the difference in area beneath the normalized $w\theta$ cospectra and the corresponding CO$_2$ and water vapor cospectra in Fig. 4.2. Thus, while using a closed-path gas analyzer eliminates the need for the WPL density corrections, alternative corrections are needed for loss of $w\chi_c$ and $w\chi_v$ covariances at high frequencies. As an example, Fig. 4.6 shows that $\lambda E$ and $F_{EC}^c$ from the open-path system (after the appropriate high-cut frequency and WPL density corrections) were 13% and 5% higher than the corresponding fluxes calculated using the closed-path analyzer. This is even after application of the theoretical transfer functions for high-frequency losses due to air flow in tubes as described in Sect. 4.1.3.1.

### 4.1.5 Sensor-Specific Corrections

#### 4.1.5.1 Flow Distortion Correction of Sonic Anemometers

Flow distortion has been a well-known problem since the beginning of sonic anemometry (Dyer 1981). The reasons are the installations of the sensors and the
Table 4.1 Classification of sonic anemometers (Based on a classification by Foken and Oncley 1995; Mauder et al. 2006)

<table>
<thead>
<tr>
<th>Anemometer class</th>
<th>Sensor type</th>
</tr>
</thead>
<tbody>
<tr>
<td>A Basic research for flux measurements</td>
<td>Kaijo-Denki A-Probe, Campbell CSAT3, Solent HS</td>
</tr>
<tr>
<td>B General use for flux measurements</td>
<td>Kaijo-Denki B-Probe, Solent Wind Master, R2, R3, METEK USA-1, Young 81000</td>
</tr>
<tr>
<td>C General use for wind measurements</td>
<td>sensors of class B, 2D-anemometer of different producers</td>
</tr>
</tbody>
</table>

size of the transmitters/receivers. For new sensors, a large ratio of the path length, $d_{pl}$, to the transmitter/receiver diameter, $a$, of up to $d_{pl}/a = 20$ is required to minimize the influence of flow distortion. Furthermore, the angle between the wind vector and the transmitter-receiver path should be large (Kaimal and Finnigan 1994). From these requirements follows a classification of sonic anemometers (Table 4.1) into those having low flow distortion but a limited open angle, which suggested for research and in omni-directional sonic anemometers for routine applications.

One approach is for manufacturers to use wind tunnel measurements to guide the design of the anemometer array in a way that reduces the amount of flow distortion and then report winds with no correction for a specified acceptance angle. This is the approach taken by Campbell Sci. with the CSAT3. Other sonic anemometer manufacturers include a flow distortion correction for horizontal flow in the firmware, unfortunately, often not described in the manual. For the sonic anemometer USA-1 and Solent HS, the correction may be turned off. These corrections not only compensate for wind tunnel measurements but also take into account experience from in situ comparisons. Here, the user has no possibility of manipulating this correction. A still open problem is whether due to this correction a self correlation between the wind components is generated which influences the flux measurements. Because of the smaller eddies close to the ground, the influence of flow distortion decreases with height. Therefore, omni-directional sonic anemometers and other sensors that do not have a large ratio between the path length and the transmitter/receiver diameter should not be used close to the ground but have no problems well above the surface.

Recently, it was proposed that this correction should also include the angle between the wind vector and the horizontal level. This correction is called angle of attack correction (van der Molen et al. 2004; Nakai et al. 2006). Anemometer-specific data are available for Solent R3 (van der Molen et al. 2004; Nakai et al. 2006); the H4 head correction for USA-1 is comparable to this correction. The application of this correction increases the fluxes significantly (Cava et al. 2008). Because all correction functions were determined in the wind tunnel, the fluxes will be overestimated and the correction should not be used (Wyngaard 1981; Högström and Smedman 2004) or used only with care.
A final approach is to use a single-path model to a three-dimensional array, incorporating the path geometry. This approach can be successful if the geometry is relatively simple, as in the ATI-K probe or the CSAT3 (van Dijk 2002). Clearly, the caveats mentioned above apply – the single-path correction must be measured in a flow with turbulence levels similar to the atmosphere.

4.1.5.2 Correction Due to Sensor Head Heating of the Open-Path Gas Analyzer LiCor 7500

The sensor head of the LiCor 7500 open-path gas analyzer is heated which can generate convection within the sampling volume and therefore (some modifications were made in LiCor 7500A) has an influence on the application of the WPL correction. Potential corrections for this effect are discussed by Burba et al. (2008), Järvi et al. (2009), and Burba and Anderson (2010). A correction using additional fine-wire thermometers within the measurement volume is described by Grelle and Burba (2007). No general consensus has been reached yet as to which method is most effective and efficient. All corrections depend on wind speed and the inclination of the sensor. Therefore, any correction should be applied with care. Overall, a correction would be larger in cold weather than in warm weather, because it is generally a function of the temperature differences between the instrument surface and the ambient air, where the instrument surface temperature is a function of thermal control, set at about 30°C, and radiation load.

Specifically, during warm weather conditions (e.g. +30°C), the difference between air and instrument surface temperatures is mostly affected by solar load, which is usually less than 1–2°C at noon. For colder ambient temperatures, the instrument surface temperature is increasingly affected by the electronics. This means that, even in summer, there is some effect due to sensor heating. However, a potential summer correction is usually smaller in absolute values, due to less need for electronics heating, and it is much smaller in relative contribution, in relation to large summer ecosystem fluxes, as compared to winter correction, which is enhanced in absolute values by strong electronics heating and in relative contribution because of very small ecosystem fluxes. A simple solution for this problem is to deploy the sensor head upside down (though still a bit inclined to allow rain to run off). In this way, the heat generated at the base of the head rises away from the sensor path. Another solution for applications where low power consumption is required could be the deployment of an enclosure around the measurement path of the LI-7500 with short tubing (Clement et al. 2009). This idea resulted in the development of the LiCor 7200 sensor, which is supposed to combine the benefits of open- and closed-path systems.

4.1.5.3 Corrections to the Krypton Hygrometer KH20

Krypton hygrometers are used to measure the water vapor content of the air by absorption of H₂O molecules in the ultraviolet spectrum. Due to the wave length
used, there is a cross sensitivity to O\textsubscript{2} molecules, which has to be corrected for as recommended by Tanner et al. (1993) and van Dijk et al. (2003).

\[
\hat{w}^f \chi^f_v = \hat{w}^f \chi^f_{v\text{KH2O}} + C_{k_o} \left( \frac{\rho_v}{\theta} \right) \hat{w}^f \theta^f
\]

where

\[
C_{k_o} = \frac{C_o m_o}{m_d} \cdot \frac{k_o}{k_v} = 0.23 \frac{k_o}{k_v} \tag{4.33}
\]

$k_o$ and $k_w$ are the KH20 extinction coefficients for oxygen and water vapor, $C_o = 0.21$ is the molar fraction of oxygen in the atmosphere, and $m_o$ is the molar mass of oxygen. The coefficients $k_w$ and $k_o$ are specific for each instrument. The extinction coefficient for water $k_w$ is given in the calibration certificate by the manufacturer. The extinction coefficient for oxygen $k_o$ can be determined experimentally. Tanner et al. (1993) recommend using a value of $k_o = -0.0045$, if the instrument-specific coefficient is not known.

### 4.1.5.4 Corrections for CH\textsubscript{4} and N\textsubscript{2}O Analyzers

In recent years, fast-response sensors for trace gases other than CO\textsubscript{2} have become commercially available. Particularly, analyzers for the greenhouse gases CH\textsubscript{4} and N\textsubscript{2}O are gaining more and more popularity in climate change research. As for the CO\textsubscript{2}/H\textsubscript{2}O analyzers, these sensors are also based on light absorption of a specific kind of molecule. However, since atmospheric concentrations of these gas species are much smaller than for CO\textsubscript{2} a more defined light source is necessary and that is why these sensors generally use lasers instead of noncoherent light. In this context, two basic measurement principles can be distinguished (Werle et al. 2008):

Tunable Diode Laser (TDL) Spectroscopy, where the output wavelength of the laser is tunable over a certain spectral range, so that the laser can scan across a specific absorption band of a molecule. For many of those analyzers, the actual measurement is carried out in an optical cell at a very low pressure. This has the disadvantage of requiring high-power pumps but leads to a wanted narrowing of the absorption bands and therefore better separation between different gas species. Examples for this type of analyzer are the Campbell TGA-100/200 or the Aerodyne QCL. The Licor LI-7700 also uses a tunable laser source but in an open-path measurement setup, similar to the LI-7500 but with a longer path length, and at ambient pressure.

Cavity Ring-Down Spectroscopy (CRDS) or Integrated Cavity Output Spectroscopy (ICOS), which generally rely upon vacuum pumps to draw down pressure in the optical cell. The intensity decay rate of light trapped in an optical cavity is measured, which is a function of the concentration of a gas species that absorbs
light at that specific wavelength. The CRDS principle is applied, for example, by
the PICARRO analyzers, whereas ICOS is used by the Los Gatos Fast Greenhouse
Gas Analyzer.

The corrections required for CH₄ and N₂O analyzers are the same as for the
commonly used H₂O or CO₂ analyzers, depending whether they have an open or
closed measurement path. Sometimes a scrubber is deployed between the tube inlet
and the measurement cell to remove water vapor from the sampling air. However,
only if all temperature and pressure fluctuations are eliminated and any humidity
is completely removed, the application of the WPL-correction is unnecessary,
(Sect. 4.1.5.4). As mentioned before, for closed-path instruments in general, a
correct determination of the time delay between the analyzer signal and the sonic
signal is crucial to obtain accurate flux estimates.

### 4.1.6 Nonrecommended Corrections

As already shown above, not all corrections are recommended for general use. The
reason is mainly that these corrections are not adequately tested or have significant
limitations. During the past 40 years several corrections have been proposed and
many are now applied, sometimes in modified versions. However, some should not
be used and these are addressed below.

Due to the presentation by Stull (1988) of the correction of humidity-dependent
fluctuations of the specific heat proposed by Brook (1978), which is some percent-
age of the flux, this correction is often used. However, shortly after the publication of
this correction several authors (Leuning and Legg1982; Nicholls and Smith 1982;
Webb 1982) showed that this correction is based on incorrect conditions, and should
never be used.

Liu et al. (2006) proposed to apply the effect of the energy balance closure by
using the WPL-correction. This is a correct assumption in the case that the energy
balance closure is based on an incorrect determination of the fluxes by the eddy
covariance method. According to the results given in Sect. 4.2, this is not the reason
for the missing energy. Therefore, this correction should not be used.

An energy balance (EB) closure adjustment of the sensible and latent heat flux
according to the Bowen and also of the CO₂ flux (Desjardins 1985; Twine et al.
2000) is not recommended. The nonclosure of the EB indicates that there is a prob-
lem, but this solution would probably be too simple, since it is unknown whether
scalar similarity can be assumed for the processes that cause an underestimation of
the eddy covariance (EC) flux. Further studies about the causes of non-closure are
warranted (see Sect. 4.2).

Furthermore, the published aliasing correction (Moore 1986) should not be
applied. Aliasing is the transformation of the energy of higher frequencies to lower
ones if the measuring system has no low-pass filters.
4.1.7 Overall Data Corrections

As shown in Sect. 4.1.2–4.1.5 most of the corrections are stability dependent or need the turbulent fluxes as input parameters. Therefore, the corrections often are performed iteratively. Though the math is tedious, Oncley et al. (2007) showed that the above corrections can be solved without iteration as a set of simultaneous equations. In Fig. 4.7, the schematic shows how to organize the system of raw data correction (Sect. 3.2.2), covariance correction (Sect. 4.1) and data quality tests as discussed in Sect. 4.3. This iteration has an effect of about 1% on the fluxes.

Furthermore, in Fig. 4.8 the effect of all correction for an about 6 week data set of the LITFASS-2003 experiment (Mengelkamp et al. 2006) is presented by Mauder and Foken (2006). Most relevant are the spectral correction (here only the shortwave part) on all fluxes and the transformation of the buoyancy flux into the sensible heat flux with a significant flux reduction. The figure shows also that the WPL correction needs a very careful application. While the latent heat flux is only slightly changed, the effect on the CO₂ flux and many other trace gas fluxes is significant (Box 4.1).

Box 4.1: Recommendations for Flux Correction

- Applying of flux correction after all raw data modifications presented in Sect. 3.2.2.
- First, all fluxes must be corrected for spectral losses (Sect. 4.1.3)

Iterate corrections until change < 0.01%

Post-field quality control (Foken et al. 2004)
  - Test for stationarity (Foken und Wichura 1996)
  - Test for integral turbulence characteristics (Foken und Wichura 1996)

Corrected and quality-assured estimates of turbulent fluxes

Fig. 4.7 Schematic of the postfield data processing (Mauder and Foken 2006)
The buoyancy flux must be transferred into the sensible heat flux, which is used in further corrections and most of the application (Sect. 4.1.2)

Water vapor and trace gas fluxes must be corrected for density fluctuations, which is different for open and closed path sensors (Sect. 4.1.4)

Several sensors need specific corrections, which may be still in development for recent sensors (Sect. 4.1.5)

The corrections should be calculated with an iterative system or a combined system of all equations (Sect. 4.1.7)

The calculation of the atmosphere-ecosystem flux needs further nonsensor-specific corrections like the storage and night time flux correction (Sect. 5.4)
4.2 Effect of the Unclosed Energy Balance

4.2.1 Reasons for the Unclosed Energy Balance

After wide application of the eddy covariance method including all corrections and after the availability of highly accurate net radiometers it became obvious that the energy balance at the Earth’s surface could not be closed with experimental data (Foken and Oncley 1995). The available energy, that is, the sum of the net radiation and the ground heat flux, was found in most cases to be larger than the sum of the turbulent fluxes of sensible and latent heat. For many field experiments and also for the CO₂ flux networks (Aubinet et al. 2000; Wilson et al. 2002), a closure of the energy balance of approximately 80% was found. The residual is

\[
\text{Res} = R_n - G - H - \lambda E, \quad (4.34)
\]

With: \( R_n \): net radiation, \( G \): soil heat flux, \( H \): sensible heat flux, and \( \lambda E \): latent heat flux (Fig. 4.9).

The problem cannot be described as only an effect of statistically distributed measuring errors because of the clear underestimation of turbulent fluxes or overestimation of the available energy. In the literature, several reasons for this incongruity have been discussed, most recently in an overview paper by Foken (2008a).

![Figure 4.9](image_url)

**Fig. 4.9** Mean diurnal cycle of all energy balance components for the maize site during the LITFASS-2003 (Mengelkamp et al. 2006) period after Liebethal (2006)
In recent papers, it was found that time-averaged fluxes (Finnigan et al. 2003) or spatially averaged fluxes including turbulent-organized structures (Kanda et al. 2004) can close the energy balance. Therefore, it must be assumed that the phenomena of the unclosed energy balance at the earth’s surface is not related to errors in the eddy covariance technique but related to atmospheric phenomena which cannot be measured with this technique. Thus, a simple correction is impossible, and how to handle this phenomena for energy fluxes and probably also for trace gas fluxes (carbon dioxide) is still an open question. Combining all findings about the problem it can be concluded that (Foken 2008a; Foken et al. 2010):

In the past, the most common point of discussion with respect to the energy balance closure problem was measurement errors, especially those of the eddy covariance technique which were assumed to cause a systematic underestimation of the turbulent fluxes. Improvements in the sensors as well as in the correction methods, and the application of a more stringent determination of the data quality have made this method much more accurate over the past 10 years (Foken et al. 2004; Moncrieff 2004; Mauder and Foken 2006; Mauder et al. 2007b). Also the analysis of the data quality of eddy covariance measurements (Mauder et al. 2006) had no remarkable effect. As shown in Sect. 4.1.7, even a careful application of all corrections of the turbulent fluxes can reduce the residual only slightly (Mauder and Foken 2006).

Different reference levels and different sampling scales of the measuring methods for net radiation, turbulent fluxes, and soil heat flux were often seen as another possible reason for the lack of energy-balance closure. Moreover, the role of energy storage in the canopy and in the soil was discussed by several authors. Most of these energy storages appear to be not significant to the problem for low vegetation canopies (Oncley et al. 2007) with the exception of the heat storage in the soil (see e.g. Culf et al. 2004; Heusinkveld et al. 2004; Meyers and Hollinger 2004; Foken 2008a).

The nonclosure of the energy balance has also been explained by the heterogeneity of the land surface (Panin et al. 1998). These authors assumed that the heterogeneity in the vicinity of a flux-measurement site generates eddies at larger time scales, but such turbulent structures generated by heterogeneities close to the measuring tower can be measured with the eddy covariance method (Thomas and Foken 2007; Zhang et al. 2007). Therefore, the low-frequency part of the spectra (Foken et al. 2006) up to about 2 h has no significant influence on the closure problem.

This problem is also closely connected with advection and fluxes associated with longer wavelengths. Some recent studies have found that fluxes averaged over long time periods of several hours (Sakai et al. 2001; Finnigan et al. 2003; Mauder and Foken 2006) or spatially averaged fluxes (Kanda et al. 2004; Inagaki et al. 2006; Steinfeld et al. 2007) could close the energy balance. During the EBEX-2000 experiment, it was found that advection can play a significant role (Oncley et al. 2007). Due to the reduction of the residual by accounting for advection, the energy balance closure problem for EBEX-2000 is smaller than similar experiments.
Fig. 4.10 Schematic figure of the generation of secondary circulations and the hypothesis of turbulent fluxes in different scales based on small eddies (s) and large eddies (l) according to Foken (2008a) where \( \langle H, E \rangle_s \) is the sensible or latent heat flux by averaging over small eddies and \( \langle H, E \rangle_l \) by averaging over large eddies. \( R_n \) is the net radiation and \( G \) is the ground heat flux.

Over homogeneous surfaces like deserts (Heusinkveld et al. 2004) or bushland (Mauder et al. 2007a) the surface energy balance can be closed. Therefore, heterogeneities on a scale larger than 100 m and up to more than 10 km currently are the focus of possible explanations.

To verify these results, area-averaged flux measurements were used during the LITFASS-2003 experiment (Beyrich and Mengelkamp 2006; Mengelkamp et al. 2006), with large aperture scintillometers (Meijninger et al. 2006), aircraft measurements, and Large-Eddy Simulations. With these area-averaging techniques a much better energy balance could be reached (Foken et al. 2010).

Combining these previous investigations, it is obvious that the correction of the unclosed energy balance cannot be a part of the eddy covariance method and its correction procedures. A conceptual picture considering the land surface–atmosphere interaction at different scales was shown by Foken (2008a). This is also based on numerical studies which have shown that at steps of heterogeneities, the fluxes are significantly larger than over more homogeneous areas (e.g. Schmid and Bünzli 1995a; Schmid and Bünzli 1995b). This was underlined by the experiments by Klaassen et al. (2002). If the size of the heterogeneities or the difference of the characteristic heterogeneities (e.g. roughness, heat fluxes) is too small, this effect disappears (Friedrich et al. 2000). In Fig. 4.10 is shown that near the surface the smaller eddies are measured with micrometeorological methods such as the eddy covariance technique and the long-wave part is not available (Steinfeld et al. 2007). The transfer of the energy from the surface to the larger eddies happens mainly at significant heterogeneities and is not uniformly distributed over the area. In the sum of the fluxes by the smaller and larger eddies, the energy can be closed as measured with area-averaging techniques and also with long-term integration. Such long-term integration shows (Fig. 4.11) that the sensible heat flux increases and closes the residual while the latent heat flux is not affected. This underlines a nonsimilarity of both fluxes which may be dependent on the transport at the heterogeneities and should be different for different sites.
4.2.2 Correction of the Unclosed Energy Balance

As shown above, the correction of the residual of the energy balance is not an issue of the correction of the eddy covariance method because the missing energy is not a missing flux at the measuring point and can be, at most, measured as advection. But such experimental setups are nearly impossible, as the EBEX-Experiment has shown (Oncley et al. 2007). If heterogeneities in the vicinity of the measuring stations are responsible, the analysis of the footprints and the footprint quality (Göckede et al. 2008) of tower sites (Chap. 8) should give a hint on this subject. But the quality of the footprint, that is, the percentages of the target area in the footprint, is not correlated with the residual. Instead, the existence of heterogeneities with a spatial scale larger than 500 m had a significant influence on the residual (Falge and Foken, 2007, personal communication). Because area-averaging flux measurements and Large Eddy Simulations are usually not available, only long-term integration (Sakai et al. 2001; Finnigan et al. 2003; Mauder and Foken 2006) can be used for the correction.

Therefore, eddy covariance measurements near the surface should not be corrected and the problem should be discussed for the lower part of the atmospheric boundary layer at spatial scales larger than the flux footprint.

As a first step, the energy exchange between the atmosphere and the underlying surface on scales larger than 1 km can be corrected with the Bowen ratio (Twine et al. 2000; Foken 2008a). According to the Bowen ratio, the residual will be distributed to the sensible and latent heat flux. This method is only valid if a similarity of both fluxes is given, which is obviously not always realized. There
are some papers available (Mauder and Foken 2006; Ingwersen et al. 2011) which propose that a large part of the unclosed energy balance is related to the sensible heat flux.

As a more realistic approach, long-term integration is proposed. About 3–5 days with similar synoptic situations should be used for each station. However, the results may have to be modified for different weather situations, wind directions, and times of the year. From the final picture comparable with Fig. 4.8, the change of the sensible and latent heat flux should be determined and used for the correction. Because secondary circulation as the possible reason for the unclosure does not occur at nighttime, only the daytime values should be corrected. On the other hand, the nighttime values of the turbulent heat fluxes are very small and each correction is within the possible statistical error. But up to now not published investigations have shown that the findings by Mauder and Foken (2006) can be site- and time-dependent.

The problem will be more complicated for trace gas fluxes like the CO$_2$ flux, because this flux cannot be corrected according to the degree of the energy balance closure as earlier proposed (Twine et al. 2000). At nighttime, no correction is necessary. For the daytime values long-term integration may be possible. But such methods are dependent on many open questions and are still in progress. A panel discussion about this subject in October 2009 could show some possible paths of research but no solution for correction (Foken et al. 2011; Box 4.2).

**Box 4.2: Recommendations Energy Balance Closure**

The phenomena of the “unclosed” energy balance in the surface layer is not a technical problem of the eddy-covariance method itself. It is related to the heterogeneous terrain and its influence on the turbulent exchange. The sensible and latent heat flux can be as a first guess corrected with the Bowen-ratio under the assumption that the scalar similarity is fulfilled. All trace gas fluxes should not be corrected.

### 4.3 Data Quality Analysis

A quality assurance (QA) and quality control (QC) process are essential for all meteorological measurements. For eddy covariance measurements, it is particularly recommended because of the very complex calculation procedure. This issue was extensively presented by Foken et al. (2004). Therefore, this chapter gives only an overview of an already published summary paper with some additional remarks.

In contrast to standard meteorological measurements (Essenwanger 1969; Smith et al. 1996; DeGaetano 1997), there are only a few papers available that discuss QC
of eddy covariance measurements (Foken and Wichura 1996; Vickers and Mahrt 1997). QC of eddy covariances should include not only tests for instrument errors and problems with the sensors but also evaluations of how closely the conditions fulfil the theoretical assumptions underlying the method. Because the latter depends on meteorological conditions, eddy covariance QC tools must be a combination of a typical test for high-resolution time series and examination of the turbulent conditions. A second problem is connected with the representativeness of the measurements depending on the footprint of the measurement. The fraction of the footprint that is in the area of interest must be calculated (Chap. 8).

Quality assurance is one of the most important issues for creation and management of a measuring program. Issues of QA are widely known for routine meteorological measuring programs (Shearman 1992). This whole book is an update of already available QA programs (e.g. Moncrieff et al. 1997; Aubinet et al. 2000; Foken et al. 2004).

The most important part of QA is QC. Several tests are discussed in this paper. QC must be done in real-time or shortly after the measurements to minimize data loss by reducing the time to detect and fix instrument problems.

### 4.3.1 Quality Control of Eddy Covariance Measurements

The QC for meteorological data follows a scheme which is similar for most of the data and is illustrated in Fig. 4.12. The first steps are automatic tests that the signal is in the typical range of the sensor. This is, in most cases, already done in the sensor software. For the raw data, several tests are necessary. The first is a check if the data are in a meteorologically possible range and the second is a set of statistical tests.

The following test is a comparison with other meteorological measurements. It is important to compare the averaged temperature, moisture, and trace gas concentrations with additional measurements. For the wind velocity, this is often not necessary.

A uniform scheme does not exist for QC of eddy covariance measurements. There is only a discussion of several aspects in the literature. In the following lines, an overview of different QC steps is given:

- The first steps of data analysis are basic tests of the raw data (Vickers and Mahrt 1997) such as automatic tests of the amplitude, the resolution of the signal, a check of the electrical and meteorological range of the data, and spikes (Højstrup 1993), which are discussed in Sect. 3.2.2.

- Statistical and uncertainty tests must be applied to sampling errors of the time series (Haugen 1978; Vickers and Mahrt 1997; Finkelstein and Sims 2001; Richardson et al. 2006) and are discussed in Chap. 7. Also steps in the time series, or reasons for nonstationarity must be identified (Mahrt 1991; Vickers and Mahrt 1997).
– A main issue for QC are tests on fulfilment of the requirements for eddy covariance measurements. Steady state conditions and a developed turbulent regime are influenced not from the sensor configuration but from the meteorological conditions (Foken and Wichura 1996). The fulfilment of these conditions is given in Sect. 4.3.2.
– A system of general quality flagging of the data is discussed in Sect. 4.3.3.
– A site-dependent QC based on footprint analysis is presented in Sect. 8.5.

4.3.2 Tests on Fulfilment of Theoretical Requirements

Foken and Wichura (1996) applied criteria for fast-response turbulence data to test for non-stationarity and substantial deviations from flux-variance similarity theory, whether due to instrumental or physical causes. The following presentation is based on Foken et al. (2004):
4.3.2.1 Steady State Tests

Steady state conditions means that all statistical parameters do not vary in time (e.g., Panofsky and Dutton 1984). Typical nonstationarity is driven by the change of meteorological variables with the time of the day, changes of weather patterns, significant mesoscale variability, or changes of the measuring point relative to the measuring events such as the phase of a gravity wave. The latter may occur because of changing footprint areas, changing internal boundary layers (especially internal thermal boundary layers in the afternoon), or by the presence of gravity waves. Presently, there are two main tests used to identify nonsteady state conditions. The first is based on the trend of a meteorological parameter over the averaging interval of the time series (Vickers and Mahrt 1997) and the second method indicates nonsteady state conditions within the averaging interval (Foken and Wichura 1996).

Vickers and Mahrt (1997) regressed the meteorological element \( s \) over the averaging interval of a time series and determined the difference of \( s \) between the beginning and the end of the time series according to this regression, \( \delta s \). With this calculation they determined the parameter of relative nonstationarity, mainly for wind components, as

\[
RN_s = \frac{\delta s}{\bar{s}}
\]

(4.35)

Measurements made over the ocean exceeded the threshold \( (RN_s > 0.50) \) 15% of the time and measurements over forest exceeded the threshold 55% of the time. A more rigorous measure of stationarity can be found in Mahrt (1998).

The steady state test used by Foken and Wichura (1996) is based on developments of Russian scientists (Gurjanov et al. 1984). It compares the statistical parameters determined for the averaging period and for short intervals within this period. For instance, the time series for the determination of the covariance of the measured signals \( w \) (vertical wind) and \( \chi s \) (horizontal wind component or scalar) of about 30 min duration will be divided into \( M = 6 \) intervals of about 5 min. \( N \) is the number of measuring points of the short interval \( (N = 6,000 \text{ for } 20 \text{ Hz scanning frequency and a } 5 \text{ min interval}) \):

\[
(w^\prime \chi_s)'_i = \frac{1}{N - 1} \left[ \sum_j w_j \cdot \chi_{sj} - \frac{1}{N} \left( \sum_j w_j \cdot \sum_j \chi_{sj} \right) \right]
\]

\[
\overline{w^\prime \chi_s}|_{SI} = \frac{1}{M} \sum_i (w^\prime \chi_s)'_i
\]

(4.36)

This value will be compared with the covariance determined for the whole interval:

\[
|\overline{w^\prime \chi_s}|_{SI} = \frac{1}{M \cdot N - 1} \left[ \sum_i \left( \sum_j w_j \cdot \chi_{sj} \right) - \frac{1}{M \cdot N} \sum_i \left( \sum_j w_j \cdot \sum_j \chi_{sj} \right) \right]
\]

(4.37)
The authors proposed that the time series is steady state if the difference between both covariances

$$RN_{C_{ov}} = \left| \frac{(w'\chi'_s)_S - (w'\chi'_s)_W}{(w'\chi'_s)_W} \right|$$

is less than 30%. This value has been found by long experience but is in good agreement with other test parameters including those of other authors (Foken and Wichura 1996).

### 4.3.2.2 Test on Developed Turbulent Conditions

Flux-variance similarity is a good measure to test the development of turbulent conditions. This similarity means that the ratio of the standard deviation of a turbulent parameter and its turbulent flux is nearly constant or a function of stability. These so-called integral turbulence characteristics are basic similarity characteristics of the atmospheric turbulence (Obukhov 1960; Wyngaard et al. 1971) and are routinely discussed in boundary layer and micrometeorology textbooks (Stull 1988; Kaimal and Finnigan 1994; Arya 2001; Foken 2008b). Foken and Wichura (1996) used functions determined by Foken et al. (1991). These functions depend on stability and have the general form for standard deviations of wind components:

$$\frac{\sigma_{u,v,w}}{u_*} = c_1 \left( \frac{h_m - d}{L} \right)^{c_2}$$

(4.39)

where $u$ is the horizontal or longitudinal wind component, $v$ the lateral wind component, $u_*$ the friction velocity, and $L$ the Obukhov length. For scalar fluxes, the standard deviations are normalized by their dynamical parameters:

$$\frac{\sigma_{\chi_s}}{\chi_s^*} = c_1 \left( \frac{h_m - d}{L} \right)^{c_2}$$

(4.40)

The constant values in Eqs. 4.39 and 4.40 are given in Table 4.2. For the neutral range, the external forcing assumed by Johansson et al. (2001) and analyzed for the integral turbulence characteristics by Thomas and Foken (2002) was considered in Table 4.3 with the latitude (Coriolis parameter $f$). The parameters given for the temperature can be assumed for most of the scalar fluxes. It must be mentioned that under nearly neutral conditions, the integral turbulence characteristics of the scalars have extremely high values (Table 4.2) and the test fails.

The test can be done for the integral turbulence characteristics of both parameters used to determine the covariance. The measured and the modeled parameters according to Eqs. 4.39 or 4.40 will be compared according to ($\chi$: $u$, $v$, $w$, or $\chi_s$)
Table 4.2 Coefficients of the integral turbulence characteristics (Foken et al. 1991, 1997; Thomas and Foken 2002)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>( (h_m-d)/L )</th>
<th>( c_1 )</th>
<th>( c_2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \sigma_w/\bar{u}_* )</td>
<td>( 0 &gt; (h_m-d)/L &gt; -0.032 )</td>
<td>1.3</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>( -0.032 &gt; (h_m-d)/L )</td>
<td>2.0</td>
<td>1/8</td>
</tr>
<tr>
<td>( \sigma_u/\bar{u}_* )</td>
<td>( 0 &gt; (h_m-d)/L &gt; -0.032 )</td>
<td>2.7</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>( -0.032 &gt; (h_m-d)/L )</td>
<td>4.15</td>
<td>1/8</td>
</tr>
<tr>
<td>( \sigma_\theta/\theta_* )</td>
<td>( 0.02 &lt; (h_m-d)/L &lt; 1 )</td>
<td>1.4</td>
<td>-1/4</td>
</tr>
<tr>
<td></td>
<td>( 0.02 &gt; (h_m-d)/L &gt; -0.062 )</td>
<td>0.5</td>
<td>-1/2</td>
</tr>
<tr>
<td></td>
<td>( -0.062 &gt; (h_m-d)/L &gt; -1 )</td>
<td>1.0</td>
<td>-1/4</td>
</tr>
<tr>
<td></td>
<td>( -1 &gt; (h_m-d)/L )</td>
<td>1.0</td>
<td>-1/3</td>
</tr>
</tbody>
</table>

Table 4.3 Coefficients of the integral turbulence characteristics for wind components under neutral conditions (Thomas and Foken 2002)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>(-0.2 &lt; (h_m-d)/L &lt; 0.4)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \sigma_w/\bar{u}_* )</td>
<td>( 0.21 \ln \left( \frac{z_+ \cdot f}{\bar{u}<em>*} \right) + 3.1; z</em>+ = 1 \text{ m} )</td>
</tr>
<tr>
<td>( \sigma_u/\bar{u}_* )</td>
<td>( 0.44 \ln \left( \frac{z_+ \cdot f}{\bar{u}<em>*} \right) + 6.3; z</em>+ = 1 \text{ m} )</td>
</tr>
</tbody>
</table>

\[
ITC_\sigma = \left| \left( \frac{\sigma_x}{\bar{x}} \right)_{\text{model}} - \left( \frac{\sigma_x}{\bar{x}} \right)_{\text{measurement}} \right| \quad (4.41)
\]

If the test parameter, integral turbulence characteristic (ITC_\sigma) is <30%, a well-developed turbulence can be assumed.

4.3.3 Overall Quality Flag System

This section also is based on the overview paper by Foken et al. (2004). The quality tests given above open the possibility to also flag the quality of a single measurement. Foken and Wichura (1996) proposed to classify the tests according to Eqs. 4.38 and 4.41 into different steps and to combine different tests. An important parameter which must be included in the classification scheme is the orientation of the sonic anemometer, if the anemometer is not an omnidirectional probe and the measuring site does not have an unlimited fetch in all directions. For these three tests, the definition of the flags is given in Table 4.4. Further tests, such as an acceptable range of the mean vertical wind velocity, can be included into this scheme.

The most important part of a flagging system is the combination of all flags into a general flag for easy use. This is done in Table 4.5 for the flags given in Table 4.4. The user of such a scheme must know the appropriate use of the flagged data. The presented scheme was classified by micrometeorological experiences so
Table 4.4 Classification of the data quality by the steady state test according to Eq. 4.38, the integral turbulence characteristics according to Eq. 4.41, and the horizontal orientation of a sonic anemometer of the type CSAT3 (Foken et al. 2004)

<table>
<thead>
<tr>
<th>Class</th>
<th>Range</th>
<th>Class</th>
<th>Range</th>
<th>Class</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0–15%</td>
<td>1</td>
<td>0–15%</td>
<td>1</td>
<td>±0–30°</td>
</tr>
<tr>
<td>2</td>
<td>16–30%</td>
<td>2</td>
<td>16–30%</td>
<td>2</td>
<td>±31–60°</td>
</tr>
<tr>
<td>3</td>
<td>31–50%</td>
<td>3</td>
<td>31–50%</td>
<td>3</td>
<td>±60–100°</td>
</tr>
<tr>
<td>4</td>
<td>51–75%</td>
<td>4</td>
<td>51–75%</td>
<td>4</td>
<td>±101–150°</td>
</tr>
<tr>
<td>5</td>
<td>76–100%</td>
<td>5</td>
<td>76–100%</td>
<td>5</td>
<td>±101–150°</td>
</tr>
<tr>
<td>6</td>
<td>101–250%</td>
<td>6</td>
<td>101–250%</td>
<td>6</td>
<td>±151–170°</td>
</tr>
<tr>
<td>7</td>
<td>251–500%</td>
<td>7</td>
<td>251–500%</td>
<td>7</td>
<td>±151–170°</td>
</tr>
<tr>
<td>8</td>
<td>501–1,000%</td>
<td>8</td>
<td>501–1,000%</td>
<td>8</td>
<td>±151–170°</td>
</tr>
<tr>
<td>9</td>
<td>&gt;1,000%</td>
<td>9</td>
<td>&gt;1,000%</td>
<td>9</td>
<td>&gt;±171°</td>
</tr>
</tbody>
</table>

Remark: The classes 1–5 for the horizontal orientation of the sonic anemometer have the same influence on the overall flagging system (Table 4.5)

Table 4.5 Proposal for the combination of the single quality flags into a flag of the general data quality (Foken et al. 2004)

<table>
<thead>
<tr>
<th>Flag of the general data quality</th>
<th>Steady state test according to Eq. 4.38</th>
<th>Integral turbulence characteristics according to Eq. 4.41</th>
<th>Horizontal orientation of the sonic anemometer</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>1–2</td>
<td>1–5</td>
</tr>
<tr>
<td>2</td>
<td>2</td>
<td>1–2</td>
<td>1–5</td>
</tr>
<tr>
<td>3</td>
<td>1–2</td>
<td>3–4</td>
<td>1–5</td>
</tr>
<tr>
<td>4</td>
<td>3–4</td>
<td>1–2</td>
<td>1–5</td>
</tr>
<tr>
<td>5</td>
<td>1–4</td>
<td>3–5</td>
<td>1–5</td>
</tr>
<tr>
<td>6</td>
<td>5</td>
<td>≤5</td>
<td>1–5</td>
</tr>
<tr>
<td>7</td>
<td>≤6</td>
<td>≤6</td>
<td>≤8</td>
</tr>
<tr>
<td>8</td>
<td>≤8</td>
<td>≤8</td>
<td>≤8</td>
</tr>
<tr>
<td>9</td>
<td>≤8</td>
<td>6–8</td>
<td>≤8</td>
</tr>
<tr>
<td></td>
<td>one flag equal to 9</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

that classes 1–3 can be used for fundamental research, such as the development of parameterizations. Classes 4–6 are available for general use such as for continuously running systems of the FLUXNET program. Classes 7 and 8 are only for orientation. Sometimes it is better to use such data instead of a gap-filling procedure, but then these data should not differ significantly from the data located before and after these data in the time series. Data of class 9 should be excluded under all circumstances. Such a scheme gives the user a good opportunity to use eddy covariance data. Finally, the data can be presented together with the quality flag as in Fig. 4.13. Most of the unusual values can be explained by the data quality flag. At night, other factors can influence the measurements. For analysis of integrated fluxes rejected
data will need to be filled in. Obviously, investigations to infer process relationships should exclude both flagged data and the gap-filled values (Box 4.3).

**Box 4.3: Recommendations or Data Quality Analysis**

The use of a data quality analysis is essential for the application of the eddy covariance technique. A physical and meteorological control of the range of the input data is insufficient. Steady state conditions and a fully developed turbulent regime are important requirements for the application of the eddy covariance technique and special tests are necessary. If the tests fail the data should be replaced with a gap-filling procedure. The application of an adequate flagging system is important for the users of the data.

### 4.4 Accuracy of Turbulent Fluxes After Correction and Quality Control

The very complex algorithm of the eddy covariance method does not allow the determination of the errors according to the error propagation law. But statistical analyses are possible to determine the uncertainties of the method (Richardson et al.)
Table 4.6 Evaluation of the accuracy of the eddy covariance method on the basis of the experimental results (Mauder et al. 2006), the data quality (Sect. 4.3), and the type of the sonic anemometer (Table 4.1, Foken and Oncley 1995)

<table>
<thead>
<tr>
<th>Sonic anemometer type</th>
<th>Data quality class</th>
<th>Sensible heat flux</th>
<th>Latent heat flux</th>
</tr>
</thead>
<tbody>
<tr>
<td>type A, e.g., CSAT3</td>
<td>1–3</td>
<td>5% or 10 Wm$^{-2}$</td>
<td>10% or 20 Wm$^{-2}$</td>
</tr>
<tr>
<td></td>
<td>4–6</td>
<td>10% or 20 Wm$^{-2}$</td>
<td>15% or 30 Wm$^{-2}$</td>
</tr>
<tr>
<td>type B, e.g., USA-1</td>
<td>1–3</td>
<td>10% or 20 Wm$^{-2}$</td>
<td>15% or 30 Wm$^{-2}$</td>
</tr>
<tr>
<td></td>
<td>4–6</td>
<td>15% or 30 Wm$^{-2}$</td>
<td>20% or 40 Wm$^{-2}$</td>
</tr>
</tbody>
</table>

2006), as described in Chap. 7. In this chapter, more empirical results are given so that the user has some hints for assessing the accuracy of the measured data.

On the basis of long-term experience in sensor comparisons and the experiments EBEX-2000 (Mauder et al. 2007b) and LITFASS-2003 (Mauder et al. 2006) and other investigations (Loescher et al. 2005) as well as software comparisons (Mauder et al. 2007b; Mauder et al. 2008), Mauder et al. (2006) have tried to give some numbers for the possible accuracy of eddy covariance measurements if they are obtained according to the present state of knowledge. A significant dependence was found on the type of sonic anemometer (Table 4.1) and on the data quality (Sect. 4.3). The results are summarized in Table 4.6. To transfer these data to the CO$_2$ flux, the results for the latent heat flux should be used with a threshold of about 0.2 mg m$^{-2}$ s$^{-1}$.

Aside from these errors, the problem of energy balance closure (Sect. 4.2) and the influence of the surrounding surface must be taken into account. The latter is discussed in relation to the footprint (Sect. 8.5). For this topic, Göckede et al. (2008) gave a classification of the measuring sites depending on the flux in the target area for different wind directions and stratifications.

Furthermore, it is important to discuss the effects of an internal boundary layer. Such a layer can exist due to a sudden change of the surface roughness or thermal conditions. The fetch from this change should be long enough that the height of the new equilibrium layer is larger than the measuring height (Stull 1988; Garratt 1990). A simple equation can be used to calculate the height of the new equilibrium layer in dependence on the fetch $x_f$ (Raabe 1991):

$$h_e = 0.3 \sqrt{x_f}$$  \hspace{1cm} (4.42)

A simple version to characterize a measuring point related to footprint and internal boundary layers can be obtained by combination of the percentage of the target area in the footprint with the height of the new equilibrium layer (Mauder et al. 2006). Fluxes should be accepted as good in quality if the percentage of the target area is higher than 80% (Göckede et al. 2008) and the new equilibrium layer is higher than the measuring height (Table 4.7).
Table 4.7  Fetch $x_f$, height of the new equilibrium layer $h_e$, and flux contribution from the target land use type dependent on the wind direction and stability for a maize site during the LITFASS-2003 experiment (Mauder et al. 2006)

<table>
<thead>
<tr>
<th></th>
<th>30°</th>
<th>60°</th>
<th>90°</th>
<th>120°</th>
<th>150°</th>
<th>180°</th>
<th>210°</th>
<th>240°</th>
<th>270°</th>
<th>300°</th>
<th>330°</th>
<th>360°</th>
</tr>
</thead>
<tbody>
<tr>
<td>$x_f$ in m</td>
<td>29</td>
<td>41</td>
<td>125</td>
<td>360</td>
<td>265</td>
<td>203</td>
<td>211</td>
<td>159</td>
<td>122</td>
<td>81</td>
<td>36</td>
<td>28</td>
</tr>
<tr>
<td>$h_e$ in m</td>
<td>1.6</td>
<td>1.9</td>
<td>3.4</td>
<td>5.7</td>
<td>4.9</td>
<td>4.3</td>
<td>4.4</td>
<td>3.8</td>
<td>3.3</td>
<td>2.7</td>
<td>1.8</td>
<td>1.6</td>
</tr>
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</table>

Flux contribution from target land use type in %

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<th>Neutral</th>
<th>Unstable</th>
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<table>
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<td>Alteddy</td>
<td>ECPack</td>
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<td>---------</td>
<td>--------</td>
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<td></td>
<td>Univ. of Bayreuth</td>
<td>Alterra</td>
<td>University of Wageningen</td>
</tr>
<tr>
<td>Data sampling</td>
<td>CSAT3, USA-1, R2, R3, T2, WMPro; CSAT3, USA-1, 6262, 7500, KH20, TGA100A, Los Gatos DLT100</td>
<td>R2, R3, CSAT3 KDTR90/TR61 7500, KH20, Lyman-α</td>
<td>R2, R3, Young; 6262, 7000, 7500, ADC OP-2</td>
</tr>
<tr>
<td>Data preparation</td>
<td>Test plausibility, spikes; Block average; optional detrending (filter); Time lag const/auto</td>
<td>Test plausibility, spikes; optional detrending (linear); Time lag const.</td>
<td>Test plausibility, spikes; Block average, optional detrending; Time lag const/auto</td>
</tr>
<tr>
<td>Coordinate rotation</td>
<td>Planar fit/2D rotation; Head-correction</td>
<td>2D rotation; Nakai et al. (2006)</td>
<td>planar fit/2D/3D rotation</td>
</tr>
</tbody>
</table>

(continued)
Table 4.8 (continued)

<table>
<thead>
<tr>
<th>Software</th>
<th>TK3</th>
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<th>EddySoft</th>
<th>EdiRE</th>
<th>eth-flux</th>
<th>TUDD</th>
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<td>Universite de Bayreuth</td>
<td>Alterra</td>
<td>University of Wageningen</td>
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<td>University of Edinburgh</td>
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<td>Technical University Dresden</td>
<td>NCAR/EOL</td>
<td>IMECC-EU/Univ. of Tuscia</td>
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<tr>
<td>hygrometer</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Moncrieff (1997), Horst (1997), Ibrom et al. (2007a); implementing Horst and Lenschow (2009)</td>
</tr>
<tr>
<td>High-frequency loss</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Webb et al. (1980), Burba et al. (2008) for closed-path; LI-COR-compliant for 7200</td>
</tr>
</tbody>
</table>
Table 4.8 (continued)

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<tr>
<th>Software</th>
<th>TK3</th>
<th>Altered</th>
<th>ECPack</th>
<th>EddySoft</th>
<th>EdiRE</th>
<th>eff-flux</th>
<th>TUDD</th>
<th>S + packages</th>
<th>ECO₂S</th>
</tr>
</thead>
<tbody>
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<td></td>
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<td>Alterra</td>
<td>University of Wageningen</td>
<td>Max-Planck-Institute Jena</td>
<td>University of Edinburgh</td>
<td>Technical University Zürich</td>
<td>Technical University Dresden</td>
<td>NCAR/EOL</td>
<td>IMECC-EU Univ. of Tuscia</td>
</tr>
<tr>
<td>Iteration of all corrections</td>
<td>yes</td>
<td>–</td>
<td>yes</td>
<td>–</td>
<td>yes</td>
<td>–</td>
<td>–</td>
<td>equations solved simultaneously, Oncley et al. (2007)</td>
<td>partial</td>
</tr>
<tr>
<td>Calculation</td>
<td>$\lambda(\theta); e_p(c_{p,day, q})$</td>
<td>$\lambda(\theta); e_p = \text{const.}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p = \text{const.}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p = \text{const.}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p = \text{const.}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p = \text{const.}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p, d_p$ and $e_p, \text{water}; \rho(\theta, p)$</td>
<td>$\lambda(\theta); e_p, d_p$ and $e_p, \text{water}; \rho(\theta, p)$</td>
<td></td>
</tr>
</tbody>
</table>

*aSimilar to EddyPro by LiCor*
4.5 Overview of Available Correction Software

The application of the different corrections is not really uniform in the applicable program packages. Nevertheless, differences between the programs are not present in basic questions but in some specific details. The comparison of the different software packages has shown (Mauder et al. 2008) that the results differ much less than the accuracy of the method. The selection of the software by the user depends much on whether the user needs an online or offline software, or software with a very fixed procedure or allowing many possibilities. But it is very important that the selection and application of the software needs micrometeorological experience. Furthermore, all constructive details which are necessary for the successful application of the software must be carefully documented during the installation of the software package. Table 4.8 gives an overview about the use of data correction and data quality testing in different applicable or widely used software packages.

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