URBAN ENERGY FLUXES IN HELSINKI AND THEIR HIGH FREQUENCY SPECTRAL CORRECTIONS

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Inadvertent climate modification has led to an increase in urban temperatures compared to the surrounding rural area. The temperature rise is argued to be caused partly by the altered energy partitioning of input net radiation to heat storage and sensible and latent heat fluxes. In addition, excess energy comes from the anthropogenic heat flux caused by traffic, residential heating and metabolism. The heat storage flux and anthropogenic heat flux have not yet been determined for any area of Helsinki and they are not directly measurable. To the contrary, turbulent fluxes of sensible and latent heat in addition to net radiation can be measured, and the anthropogenic heat flux together with the heat storage flux can be solved as a residual. As a result, all inaccuracies in the determination of the energy balance components propagate to the residual term and special attention must be paid to the accurate determination of the components. One cause of error in the turbulent fluxes is the fluctuation attenuation at high frequencies which can be accounted for by high frequency spectral corrections.

The aim of this study is twofold: to assess the relevance of high frequency corrections to water vapor fluxes and to assess the temporal variation of the energy fluxes. Turbulent fluxes of sensible and latent heat have been measured at SMEAR III station, Helsinki, since December 2005 using the eddy covariance technique. In addition, net radiation measurements have been ongoing since July 2007. The used calculation methods in this study consist of widely accepted eddy covariance data post processing methods in addition to Fourier and wavelet analysis.

The high frequency spectral correction using the traditional transfer function method is highly dependent on relative humidity and has an 11% effect on the latent heat flux. This method is based on an assumption of spectral similarity which is shown not to be valid. A new correction method using wavelet analysis is thus developed and tested and it seems to account for the high frequency variation deficit. Anyhow, the resulting wavelet correction remains minimal in contrast to the traditional transfer function correction. The energy fluxes exhibit a behavior characteristic for urban environments: the energy input is channeled to sensible heat as latent heat flux is restricted by water availability. The monthly mean residual of the energy balance ranges from 30 Wm$^{-2}$ in summer to -35 Wm$^{-2}$ in winter meaning a heat storage to the ground during summer. Furthermore, the anthropogenic heat flux is approximated to be 50 Wm$^{-2}$ during winter when residential heating is important.
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## List of variables

A list of variables introduced in the text is represented below. The abbreviations in "Usage" are as follows: DFT – discrete Fourier transform, DWT – discrete wavelet transform, TFC – transfer function correction, ST – similarity theory, SF – structure function.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Explanation</th>
<th>Usage</th>
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<tbody>
<tr>
<td>$A$</td>
<td>mean of variable $A$</td>
<td>DFT</td>
</tr>
<tr>
<td>$A_m$</td>
<td>amplitude at index $m$</td>
<td>DFT</td>
</tr>
<tr>
<td>$\alpha_u$</td>
<td>Kolmogorov constant for one-dimensional $u$ spectra (=0.55)</td>
<td>ST</td>
</tr>
<tr>
<td>$\beta_1$</td>
<td>Kolmogorov constant for one-dimensional $\theta$ spectra (=0.8)</td>
<td>ST</td>
</tr>
<tr>
<td>$c_p$</td>
<td>specific heat of air at constant pressure (=1004 Jkg$^{-1}$K$^{-1}$)</td>
<td>fluxes</td>
</tr>
<tr>
<td>$C, C(j)$</td>
<td>co-spectrum, co-spectrum at scale index $j$</td>
<td>DFT, DWT</td>
</tr>
<tr>
<td>$CF$</td>
<td>correction factor ($= \overline{w's'_\text{meas}/w's'}$)</td>
<td>TFC</td>
</tr>
<tr>
<td>$D_{assu}$</td>
<td>mixed velocity scalar structure function</td>
<td>SF</td>
</tr>
<tr>
<td>$\Delta k_j$</td>
<td>change in wave number</td>
<td>DWT</td>
</tr>
<tr>
<td>$\delta_{ij}$</td>
<td>Kronecker delta</td>
<td></td>
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<tr>
<td>$d_y$</td>
<td>distance ($= U/f_s$)</td>
<td>DWT</td>
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<td>$\tau, e'$</td>
<td>turbulent kinetic energy per mass unit [m$^2$s$^{-3}$], mean and deviation</td>
<td></td>
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<tr>
<td>$E(k_1)$</td>
<td>one-dimensional spectral energy density</td>
<td>ST</td>
</tr>
<tr>
<td>$\epsilon$</td>
<td>dissipation rate [m$^2$s$^{-3}$]</td>
<td>ST</td>
</tr>
<tr>
<td>$\epsilon_{EB}$</td>
<td>error term in the energy balance</td>
<td>fluxes</td>
</tr>
<tr>
<td>$\varepsilon$</td>
<td>emissivity</td>
<td>fluxes</td>
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<tr>
<td>$f$</td>
<td>dimensionless frequency $= n(z - z_d)/U$</td>
<td>spectra</td>
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<tr>
<td>$f_m$</td>
<td>peak dimensionless frequency of a power or co-spectrum</td>
<td>spectra</td>
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<tr>
<td>$f(t)$</td>
<td>data record</td>
<td>DWT</td>
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<tr>
<td>$f_s$</td>
<td>sampling frequency (10 Hz for SMEAR III)</td>
<td>spectra</td>
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<td>$F$</td>
<td>flux</td>
<td>footprint</td>
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<tr>
<td>$F_A, F_{Ar}, F_{Ai}$</td>
<td>Fourier transform amplitudes of variable $A$, real and imaginary parts</td>
<td>DFT</td>
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<tr>
<td>$H$</td>
<td>sensible heat flux [Wm$^{-2}$]</td>
<td>fluxes</td>
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<tr>
<td>$j$</td>
<td>scale index $[1, J]$</td>
<td>DWT</td>
</tr>
<tr>
<td>$J$</td>
<td>data record length $N = 2^J$</td>
<td>DWT</td>
</tr>
<tr>
<td>$k$</td>
<td>wave number (= $2\pi n/U$)</td>
<td>spectra</td>
</tr>
<tr>
<td>$k_1$</td>
<td>one dimensional wave number (x-direction)</td>
<td>ST</td>
</tr>
<tr>
<td>$\kappa_v$</td>
<td>von Kármán constant (=0.4)</td>
<td>ST</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Unit(s)</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td>---------</td>
</tr>
<tr>
<td>$L$</td>
<td>Obukhov length</td>
<td>m</td>
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<td>$L_v$</td>
<td>latent heat of vaporization</td>
<td>fluxes</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>wave length</td>
<td></td>
</tr>
<tr>
<td>$LE$</td>
<td>latent heat flux $[\text{Wm}^{-2}]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$n$</td>
<td>natural frequency $[\text{Hz}]$</td>
<td>spectra</td>
</tr>
<tr>
<td>$n_m$</td>
<td>peak frequency of power or co-spectrum $[\text{Hz}]$</td>
<td>spectra</td>
</tr>
<tr>
<td>$n_N$</td>
<td>frequency, number of cycles per data period</td>
<td>DFT</td>
</tr>
<tr>
<td>$N$</td>
<td>data record length</td>
<td>DFT, DWT</td>
</tr>
<tr>
<td>$N_s$</td>
<td>half of the scalar dissipation rate analogous to $\epsilon$</td>
<td>SF</td>
</tr>
<tr>
<td>$p'$</td>
<td>pressure fluctuation</td>
<td></td>
</tr>
<tr>
<td>$p$</td>
<td>position index</td>
<td>DWT</td>
</tr>
<tr>
<td>$\phi$</td>
<td>footprint function</td>
<td>footprint</td>
</tr>
<tr>
<td>$\psi$</td>
<td>wavelet function</td>
<td>DWT</td>
</tr>
<tr>
<td>$\phi_e$</td>
<td>dimensionless dissipation rate of energy</td>
<td>ST</td>
</tr>
<tr>
<td>$\phi_N$</td>
<td>dimensionless dissipation rate of heat</td>
<td>ST</td>
</tr>
<tr>
<td>$\phi_h$</td>
<td>dimensionless gradient of heat</td>
<td>ST</td>
</tr>
<tr>
<td>$\phi_m$</td>
<td>dimensionless wind shear</td>
<td>ST</td>
</tr>
<tr>
<td>$\phi_q$</td>
<td>dimensionless gradient of humidity</td>
<td>ST</td>
</tr>
<tr>
<td>$q, q'$</td>
<td>water vapor concentration and its deviation from the mean</td>
<td>fluxes</td>
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<tr>
<td>$q_s$</td>
<td>scaling humidity</td>
<td>ST</td>
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<tr>
<td>$Q$</td>
<td>quadrature spectrum</td>
<td>spectra</td>
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<tr>
<td>$Q(\hat{x})$</td>
<td>sink/source strength of surface vegetation volume</td>
<td>footprint</td>
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<td>$Q_A$</td>
<td>energy advection $[\text{Wm}^{-2}]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$Q_F$</td>
<td>anthropogenic heat flux $[\text{Wm}^{-2}]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$Q_S$</td>
<td>storage heat flux $[\text{Wm}^{-2}]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$r$</td>
<td>separation distance between two points</td>
<td>SF</td>
</tr>
<tr>
<td>$R$</td>
<td>correlation between two variables</td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>net radiation $[\text{Wm}^{-2}]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$R_j$</td>
<td>spatial scale at scale index $j = 2^j dy$</td>
<td>DWT</td>
</tr>
<tr>
<td>$R_i_f$</td>
<td>flux Richardson’s number</td>
<td>ST</td>
</tr>
<tr>
<td>$Res$</td>
<td>residual of the energy balance $= R_n - H - LE$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$RH$</td>
<td>relative humidity $[%]$</td>
<td>fluxes</td>
</tr>
<tr>
<td>$\rho$</td>
<td>air density $[\text{kgm}^{-3}]$</td>
<td></td>
</tr>
<tr>
<td>$s$</td>
<td>scalar</td>
<td>SF</td>
</tr>
<tr>
<td>$s$</td>
<td>scaling parameter</td>
<td>DWT</td>
</tr>
<tr>
<td>$X_s$</td>
<td>$s$ denotes a scalar when used as a subscript</td>
<td></td>
</tr>
<tr>
<td>$S, S(j)$</td>
<td>spectral energy density, spectral energy density at scale index $j$</td>
<td>DFT, DWT</td>
</tr>
<tr>
<td>Symbol</td>
<td>Definition</td>
<td>Unit</td>
</tr>
<tr>
<td>--------</td>
<td>------------------------------------------------</td>
<td>---------------</td>
</tr>
<tr>
<td>$SN$</td>
<td>signal-to-noise ratio</td>
<td>DWT</td>
</tr>
<tr>
<td>$S_{D_{ssu}}$</td>
<td>slope of $D_{ssu}$ ($= d\log(D_{ssu})/d\log(r)$)</td>
<td>SF</td>
</tr>
<tr>
<td>$\sigma^2_A$</td>
<td>variance of variable $A$</td>
<td>DFT</td>
</tr>
<tr>
<td>$\sigma_s$</td>
<td>standard deviation of the time series of a scalar</td>
<td>DWT</td>
</tr>
<tr>
<td>$\sigma^{(j)}$</td>
<td>standard deviation of power spectral density</td>
<td>DWT</td>
</tr>
<tr>
<td>$t$</td>
<td>time in $f(t)$</td>
<td>DWT</td>
</tr>
<tr>
<td>$t_r$</td>
<td>translation parameter</td>
<td>DWT</td>
</tr>
<tr>
<td>$T$</td>
<td>averaging period</td>
<td>TFC</td>
</tr>
<tr>
<td>$T, T'$</td>
<td>temperature and its deviation from the mean [K]</td>
<td>fluxes ST</td>
</tr>
<tr>
<td>$T_*$</td>
<td>scaling temperature</td>
<td>TFC</td>
</tr>
<tr>
<td>$TF, TF_{HF}, TF_{LF}$</td>
<td>transfer function, $TF$ high frequency, $TF$ low frequency</td>
<td>TFC</td>
</tr>
<tr>
<td>$\tau$</td>
<td>momentum stress [kgms$^{-2}$]</td>
<td>fluxes ST</td>
</tr>
<tr>
<td>$\tau_s$</td>
<td>response time of a scalar [s]</td>
<td>TFC</td>
</tr>
<tr>
<td>$\theta, \theta'$</td>
<td>potential temperature and its deviation from the mean [K]</td>
<td>TFC</td>
</tr>
<tr>
<td>$\theta_v, \theta'_v$</td>
<td>virtual potential temperature and its deviation from the mean [K]</td>
<td>TFC</td>
</tr>
<tr>
<td>$u, u'$</td>
<td>wind speed in x-direction and its deviation from the mean</td>
<td>ST</td>
</tr>
<tr>
<td>$\bar{u}, U$</td>
<td>mean horizontal wind speed</td>
<td>ST</td>
</tr>
<tr>
<td>$\pi_j$</td>
<td>three components of mean wind speed ($j=1,2,3$)</td>
<td>DWT</td>
</tr>
<tr>
<td>$u_*$</td>
<td>friction velocity</td>
<td></td>
</tr>
<tr>
<td>$v, v'$</td>
<td>wind speed in y-direction and its deviation from the mean</td>
<td></td>
</tr>
<tr>
<td>$w, w'$</td>
<td>wind speed in z-direction and its deviation from the mean</td>
<td></td>
</tr>
<tr>
<td>$Wf^{(j)}[p]$</td>
<td>wavelet coefficient at scale index $j$ and position index $p$</td>
<td>DWT</td>
</tr>
<tr>
<td>$WN$</td>
<td>maximum amplitude of white noise</td>
<td></td>
</tr>
<tr>
<td>$x_j$</td>
<td>$x, y$ and $z$ directions ($j=1,2,3$)</td>
<td></td>
</tr>
<tr>
<td>$z$</td>
<td>height above ground [m], also measurement height</td>
<td></td>
</tr>
<tr>
<td>$z_0$</td>
<td>roughness length [m]</td>
<td></td>
</tr>
<tr>
<td>$z_d$</td>
<td>displacement height [m]</td>
<td></td>
</tr>
<tr>
<td>$z_H$</td>
<td>mean building height [m]</td>
<td></td>
</tr>
<tr>
<td>$z_r$</td>
<td>roughness sublayer height [m]</td>
<td></td>
</tr>
<tr>
<td>$\zeta$</td>
<td>stability ($= (z - z_d)/L$)</td>
<td>ST</td>
</tr>
</tbody>
</table>
Introduction

To date, the ever-increasing population of the world is 6.9 billion from which about half are city dwellers (figure 1). According to the prediction of the United Nations, 70% of the people will be living in urban surroundings in 2050. This development will most surely alter the urban climate through changes in land use and the atmosphere. In essence, the inadvertent climate modification stems from the modification of the exchange of energy and momentum between the atmosphere and the city. The fluxes are characterized by a high degree of heterogeneity which makes urban environments especially challenging to study. Cities comprise of a wide variety of surface types ranging from impervious surfaces, such as parking lots, through rough elements, such as buildings, to vegetation. Urban environments also differ from city to city making their investigation ever demanding and leading to a need of extensive urban research covering a range of city types.

The recognition of the need of urban research has led to a selection of studies mainly conducted in developed countries. Grimmond and Oke (2002) give an overview of several studies in North America. Also studies in European cities are

Figure 1.1: The development of the world population. Data from the United Nations Population Division (http://esa.un.org/unpp).
numerous (for example, Christen and Vogt (2004), Basel, Switzerland; Grimmond et al. (2004), Marseille, France; Offerle et al. (2005), Łódź, Poland; Vesala et al. (2008), Helsinki, Finland; Pigeon et al. (2008), Toulouse, France). To the contrary, only a few studies of Asian cities have been so far completed (e.g. Moriwaki and Kanda (2006), Tokyo, Japan and Al-Jiboori et al. (2002), Beijing, China). In spite of the grown research interest, urban studies remain scarce in comparison with vegetated surfaces such as forests or agricultural areas.

This work is based on data from a multidisciplinary urban micrometeorological study site (SMEAR III) that was established in Helsinki in 2004. The measurements include continuous state of the art flux measurements of sensible and latent heat with the eddy covariance technique (EC). The measurements consist of a fast response sonic anemometer and two different types of infra-red gas analyzers (IRGA). In addition, data on air composition and meteorological variables are available (Järvi et al., 2009a).

The objective of this work is twofold. First, the aim is to study the temporal variability in energy fluxes having the emphasis on the heat storage to the ground and on the anthropogenic heat flux. These terms cannot be directly measured and are thus solved as a residual of the energy balance. Consequently, all errors in the energy balance terms tend to propagate to the determination of the anthropogenic heat flux and the heat storage flux, and special attention must be paid to any small corrections in the turbulent heat fluxes. One cause of error in these fluxes is the high frequency variance deficit caused by the incapability of the measurement system to solve small scale variation. Hence, the second aim of this work is to assess the importance of high frequency spectral corrections on urban fluxes.

In section 2, the theory and background of the study will be presented. This includes fundamental concepts of the atmospheric boundary layer, basics of time series analysis, boundary layer similarity theories and turbulence spectra. In addition, the urban boundary layer is discussed as a special case of a boundary layer. In section 3, the measurement site, data acquisition and post processing are presented. In addition, it is shown how the spectral corrections are conducted with the widely used transfer function method, and a new wavelet-based correction is developed. In section 4, complementary meteorological data is shown, spectral similarity is discussed and the results of the two spectral correction methods are shown. Furthermore, the urban energy fluxes are discussed starting with net radiation and continuing with
the sensible and latent heat fluxes. Finally, the seasonal, diurnal and wind direction
dependent variation of the residual of the energy balance is shown. An estimate for
the anthropogenic heat flux is given and the results are compared with the values
from a literature review. In section 5, the study is summarized and final conclusions
are given.
In this section, the definitions of the planetary boundary layer, turbulence, turbulent fluxes and stability are gone through. Basics of time series analysis are represented briefly and the analysis methods are used for determining turbulent spectra. Finally, an overview of the urban boundary layer as a special case of a boundary layer is represented.

2.1 Planetary boundary layer

The planetary boundary layer (PBL, a.k.a. atmospheric boundary layer) is the lowest part of the atmosphere reaching up to a few tens of meters in stable conditions up to a few kilometers in unstable conditions. In this layer, the flow is always turbulent and the air mass is directly influenced by the vicinity of the surface; Momentum, heat and water vapor fluxes are the means of communication between the atmosphere and the surface. According to one definition, the planetary boundary layer is the air volume where a change at the surface alters the conditions within an hour. Such a change might be for instance an abrupt release of air pollution (Högström and Smedman 1989). The PBL is of special interest to the human kind since most of us live in the PBL.

The planetary boundary layer can be further divided into a surface layer near to the ground and an Ekman layer lying on top. The surface layer is characterized by sharp vertical gradients of such properties as moisture and temperature, and horizontal wind speed follows a logarithmic law in neutral conditions. The turbulent fluxes stay fairly constant within the layer which is extremely important for the
applicability of micrometeorological measurement methods. In the Ekman layer, on the other hand, fluxes decrease with height and the wind speed increases and turns in a spiral with height. (Holton, 2004)

The remaining part of the atmosphere above the planetary boundary layer is called the free atmosphere. Here the wind speed has reached its geostrophic value and is constant, to a certain extent, with height. Turbulence can be ignored in an approximate treatment of synoptic-scale flows (Holton, 2004).

2.2 Turbulence, turbulent fluxes and atmospheric stability in the surface layer

2.2.1 Turbulence

Turbulence consists of 3-dimensional, chaotic, irregular swirls of motion superimposed on a mean flow. These eddies have different sizes and they are also superimposed on each other. The relative strengths of these eddies define the turbulence spectrum (section 2.4). The eddies can be characterized either by their period ($T$) or dimension ($\lambda$) and the transition between them is done with Taylor’s hypothesis of frozen turbulence $T = \lambda/U$, where $U$ is the mean wind speed (Taylor, 1938). Here the turbulence is assumed to advect past a sensor without changing, i.e. frozen (Stull, 1988).

A flow can either be turbulent or laminar. In a laminar case, the streamlines of a fluid are fairly smooth. The flow type is defined by the Reynolds’ number ($Re$, see Table 3.4.1) which is a dimensionless number showing the ratio between inertial and viscous forces in a fluid. When $Re$ is large, the flow is considered turbulent. In the atmosphere, a laminar layer with a height of a couple of millimeters can be found close to the ground, but above it the atmosphere is almost always turbulent.

Turbulence in the atmosphere is produced by two mechanisms: thermally and mechanically. Thermal production occurs when the atmosphere is unstably stratified (section 2.2.4) and warm air rises as a result of buoyancy. Mechanical production is induced either by a wind share or obstacles. When air flow encounters an obstacle, e.g. a building, turbulence is caused adjacent to and downwind of the obstacle (section 2.5.1). Shear-induced turbulence is always present near the ground due
to the no-slip boundary at the surface, i.e. the wind speed goes to zero. It is well known that turbulent transport dominates by far over molecular diffusion but viscosity has an important role because of this no-slip boundary; even weak winds cause a significant velocity shear at the surface leading to a transport of momentum. (Stull [1988])

A common way to study turbulence is to split the variable of interest (wind speed, temperature, water vapor) to a mean part and a fluctuating part according to Reynolds’ decomposition (Reynolds [1894])

\[ X = \bar{X} + X' \]  \hspace{1cm} (2.1)

Here \( X \) is a variable, \( \bar{X} \) is its mean value and \( X' \) the perturbation from the mean. The variance of a term is defined with this notation as \( \sigma^2_X = \bar{X'}^2 \) and covariance between two variables as \( \text{cov}_{X_1, X_2} = \bar{X'_1 X'_2} \). According to definition, \( \bar{X'} = 0 \).

### 2.2.2 Turbulent fluxes

As stated before, turbulence is an effective way of transporting heat, momentum and scalars in the atmospheric boundary layer. The turbulent fluxes of latent \( (LE) \) and sensible heat \( (H) \) in addition to momentum stress \( (\tau) \) are defined as

\[
\begin{align*}
H &= \rho c_p w' \theta'_v \quad \text{sensible heat} \\
LE &= L_v \rho w' q' \quad \text{latent heat} \\
\tau &= -\rho w' u' \quad \text{momentum},
\end{align*}
\]  \hspace{1cm} (2.2)

where \( w' \), \( \theta'_v \), \( q' \) and \( u' \) are the deviations from average vertical wind speed, potential virtual temperature \( ^1 \) water vapor concentration and horizontal wind speed, respectively (e.g. Stull [1988]). \( \rho \) is the density of air, \( c_p \) the specific heat of air at constant pressure and \( L_v \) is the latent heat of vaporization of water. The potential temperature can be approximated by normal temperature if the measurements are done close to the ground. The fluxes of carbon dioxide and particles can be defined in a similar way. The fluxes of heat and water vapor are positive when directed upwards indicating a positive correlation between \( w' \) and \( T' \) or \( q' \). \( \tau \) is defined positive when the correlation between \( w' \) and \( u' \) is negative, which is usually the case as

\( ^1 \)Potential temperature: \( \theta = T(1000/p)^{R/c_p} \); \( p=\text{pressure}, \; R=287 \; \text{Jkg}^{-1}\text{K}^{-1}, \; c_p=1004 \; \text{Jkg}^{-1}\text{K}^{-1} \); Virtual temperature \( T_v = T(1 + 0.61q) \); Virtual potential temperature: \( \theta_v = (\theta/T) \cdot T_v \) (Kaimal and Finnigan [1994]).
wind speed increases with height and momentum is transported down the gradient. The eddy covariance method is based on the measurement of fluxes according to equations 2.2. The scaling variables needed below (section 2.4) can also be defined using covariances

\[
\begin{align*}
    u_* &= \sqrt{-w'u'} & \text{friction velocity} \\
    T_* &= -\frac{w'q'}{u_*} & \text{scaling temperature} \\
    q_* &= -\frac{w'q'}{u_*} & \text{scaling humidity.}
\end{align*}
\]

(2.3)

2.2.3 Flux footprint area

When turbulent fluxes are measured at a certain point from a tower, the sensor does not detect the fluxes just below the tower but from an ellipsoidal area upwind from the tower (Vesala et al., 2008). This area is called the flux footprint and it is a complex function of the roughness length, wind speed, stability and measurement height. In unstable conditions the signal originates from near the sensor and, as an opposite, in very stable conditions the signal comes from a longer distance away. Furthermore, in windy conditions turbulent production by shear is larger and thus the signal comes from near the tower whereas in calm conditions the shear production is weak and the flux originates from further away. The source area also depends on the height of the surface studied: the source area of a canopy is much smaller than that of a surface as turbulence at the surface is weaker.

The footprint function, \( \varphi \), describes how much a certain area contributes to the flux observed by the sensor. When this function is multiplied by a function telling the source/sink strength of a surface-vegetation volume, \( Q(\hat{x}) \), and the product is further integrated in space, the result is the flux (Vesala et al., 2008)

\[
F = \int \varphi(x, \hat{x})Q(\hat{x})d\hat{x}.
\]

(2.4)

Here \( x \) is a vector indicating the location of the sensor. The integration is done from \(-\infty\) to \(\infty\) both in \( x \)- and \( y \)-directions. The footprint function extends to infinity and hence usually only the area corresponding to a certain percentage (often 80%) of the flux contribution is given. Figure 2.1 is a schematic representation of the concept of a footprint area and footprint function.
2.2.4 Turbulent kinetic energy and stability

The turbulent kinetic energy (TKE) per mass unit can be derived from the variances of the three wind components as

$$e = \frac{1}{2} \left( w'^2 + v'^2 + w'^2 \right).$$ \hspace{1cm} (2.5)

The rate of change of the turbulent kinetic energy with time is

$$\frac{\partial e}{\partial t} = - \sum_{j} \frac{\partial e}{\partial x_j} \partial_{ij} \left( u_i' \theta_v' \right) - \sum_{j} \left( u_j' u_j' \frac{\partial e}{\partial x_j} - \frac{\partial (u_j' e)}{\partial x_j} \right) - \frac{1}{\rho} \frac{\partial (\rho u_i' p')}{\partial x_i} - \epsilon. \hspace{1cm} (2.6)$$

Here $u_j'$ and $u_i'$ represent both the three wind components, $\delta$ is the Kronecker delta, $g$ the gravitational acceleration on Earth and $\theta_v$ the virtual potential temperature. V I is the advection of TKE and the term can be neglected when horizontal homogeneity is assumed. II is buoyant production or loss, the conversion between mean flow potential energy and TKE. It is positive when the motions lower the center of mass, that is, in unstable situations when warm and light air rises. In such a case convective eddies have positively correlated $w'$ and $T'$. When the atmosphere
is stably stratified, this term acts to suppress turbulence. Term III corresponds to the shear production of turbulence where energy is converted from the mean flow to TKE. It is always positive independent of stability. Term IV (turbulent transport term) is the rate by which turbulence is imported or exported due to velocity fluctuations. Term V (dissipation term) corresponds to pressure transport and it cannot be accurately measured. Term VI is the rate at which turbulent energy is dissipated into internal energy and it can be calculated from the high frequency end of the turbulence spectrum (section 2.2.4) (Kaimal and Finnigan, 1994). Stability can be assessed with the flux Richardsons number which is the ratio between buoyant production or loss and shear production (Richardson, 1920)

\[ \frac{\text{term II}}{\text{term III}} = -\frac{z\kappa_v \frac{g}{\rho_v} u'\theta'_v}{u'^3} \]  

(2.7)

Here the relation \( \frac{du}{dz} = \frac{u_*}{z\kappa_v} \) was used and \( \kappa_v = 0.4 \) is the von Kármán constant. If \( Ri_f < 0 \), the layer is unstably stratified and turbulence is present as both buoyancy and wind shear act to increase turbulence. If the air is stably stratified \( Ri_f > 0 \) and turbulence occurs only if the shear production is strong enough to keep \( Ri < 0 \).

Another variable called the Obukhov length (Obukhov, 1946) is related to \( Ri_f \) as

\[ L = \frac{z}{Ri_f} = -\frac{u'^3}{\kappa_v \frac{g}{\rho_v} u'\theta'_v} = -\frac{\rho c_p u'^3}{\kappa_v \frac{g}{\rho_v} H} \]  

(2.8)

The Obukhov length with units of meters does not vary with height but time. The atmosphere is neutrally stratified when \( L \rightarrow \pm \infty \). Stable stratification occurs when \( L > 0 \) and unstable when \( L < 0 \) (\( u'\theta'_v > 0 \)). The Obukhov length can also be interpreted as the height where mechanical and thermal production are equal. When \( z < |L| \) mechanical production dominates and when \( z > |L| \) thermal dominates. The Obukhov length is the key parameter in the Monin-Obukhov similarity theory introduced in section 2.4.

### 2.3 Basics of time series analysis

#### 2.3.1 Discrete Fourier transform

In order to represent a time series as a spectrum, one must convert the time series into a function of frequency. A commonly used way to do this is to make a Fourier
series by making a Fourier transform. To do this, the time series must be well-behaved and continuous. The discrete Fourier transform (DFT) of variable \( A \) is defined as

\[
F_A(n_N) = \frac{1}{N} \sum_{m=0}^{N-1} A(m) \cos \left( \frac{2\pi n_N m}{N} \right) - i \frac{1}{N} \sum_{m=0}^{N-1} A(m) \sin \left( \frac{2\pi n_N m}{N} \right).
\]  

(2.9)

Here \( F_A(n_N) \) is an imaginary number, which consists of a real part representing the amplitudes \( A(m) \) of sine waves and an imaginary part corresponding to the amplitudes of cosine waves. \( N \) is the number of data points and \( n_N \) is the frequency as the number of cycles per data period. The frequency in Hz \( n = n_N f_s \), where \( f_s \) is the sampling frequency. The variance is defined

\[
\sigma_A^2 = \frac{1}{N} \sum_{m=0}^{N-1} (A_m - \overline{A})^2 = \sum_{n_N=1}^{N-1} |F_A(n_N)|^2 = \sum_{n_N=1}^{N-1} F_{Ar}(n_N)^2 + F_{Ai}(n_N)^2,
\]

(2.10)

where \( F_{Ar} \) and \( F_{Ai} \) are the real and imaginary parts of \( F_A \), respectively.

The spectral energy of a variable \( A \) at frequency \( n \) can be defined as

\[
G_A(n_N) = F_A^*(n_N) \cdot F_A(n_N) = F_{Ar}(n_N)^2 + F_{Ai}(n_N)^2 = |F_A(n_N)|^2,
\]

(2.11)

where \( F_A^* \) is the complex conjugate of \( F_A \). The Spectral energy density \( S(n_N) \), on the other hand, gives the fraction of variance of a certain variable at a certain frequency interval \( \Delta n_N \)

\[
S_A(n_N) = \frac{2|F_A(n_N)|^2}{\Delta n_N}.
\]

(2.12)

A plot of \( S_A(n_N) \) against \( n_N \) is called the power spectrum. The integral of \( S(n_N) \) over \( n_N \) yields the total variance of the variable, as should be expected (e.g. Stull (1988))

\[
\sigma_A^2 = \int_{n_N} S_A(n_N) dn_N.
\]

(2.13)

Similarly to equation (2.11), the combined spectral intensity of two variables \( A \) and \( B \), that is the cross spectrum is defined as

\[
G_{AB}(n) = F_A^*(n) \cdot F_B(n)
\]

(2.14)

\[
= C + iQ
\]

(2.15)

\[
C = F_{Ar} F_{Br} + F_{Ai} F_{Bi}
\]

(2.16)

\[
Q = F_{Ai} F_{Br} - F_{Ar} F_{Bi},
\]

(2.17)
where $C$ is the co-spectrum and $Q$ the quadrature spectrum. The densities of $C$ and $Q$ corresponding to equation 2.12 are also often used. The quadrature spectrum equals the spectrum of the product of $B$ times $A$, that is phase shifted by $\pi/2$ and is thus not used often in micrometeorology. The co-spectrum, on the other hand, is used in the following because the sum of all $C(n)$ equals the covariance between variables $A$ and $B$

$$AB^* = \int C_{A,B}(n)dn,$$  

(2.18)

where $C_{A,B}(n)$ is now the co-spectral density.

### 2.3.2 Discrete wavelet transform

The discrete Fourier transform gives information only on the frequency of perturbations, not on the location of these frequencies in time domain. To the contrary, wavelet transform can be used for gaining information on the detailed structure of turbulent transfer as it preserves information on both scale and time. This time-localization was one of the original motives in the development of wavelet analysis in the 80s (Mallat, 1999). Another positive feature of the wavelet transform is that it can be applied to non-continuous data as an opposite to DFT.

Instead of extended sine and cosine waves, like in DFT (equation 2.9), wavelet analysis uses localized pulses called wavelets (a.k.a. basis function, time-frequency atom). A wavelet is a function $\psi \in L^2(\mathbb{R})$ with a zero average $\int_{-\infty}^{\infty} \psi(t)dt = 0$. It is also normalized so that $||\psi|| = 1$. A family of wavelets is obtained by scaling $\psi$ by $s$ and translating it by $t_r$

$$\psi_{t_r,s}(t) = \frac{1}{\sqrt{s}} \psi \left( \frac{t - t_r}{s} \right).$$  

(2.19)

Here $s$ is the dilation parameter which gives the time scale, that is, the duration and the amplitude of the wavelet. $t_r$ gives the center of the wavelet. The first and perhaps the simplest wavelet is the Haar wavelet (figure 2.2) which is defined

$$\psi = \begin{cases} 1 & 0 \leq t_r < 0.5 \\ -1 & 0.5 \leq t_r < 1 \\ 0 & \text{otherwise.} \end{cases}$$  

(2.20)

(2.21)
Figure 2.2: Examples of two wavelets: the Haar and the Mexican hat wavelet.

Figure 2.2 also shows an example of one other wavelet, the Mexican hat (Witkin, 1983).

The wavelet transform of function $f \in L^2(\mathbb{R})$ at time $t_r$ and scale $s$ is

$$Wf(t_r, s) = \int_{-\infty}^{\infty} f(t) \frac{1}{\sqrt{s}} \psi\left(\frac{t - t_r}{s}\right) dt.$$ (2.22)

This convolution measures the variation of $f$ in the neighborhood of $t_r$ with a scale proportional to $s$. In a discrete wavelet transform (DWT), $t_r$ and $s$ are discrete, and if dyadic scales are used for the sake of computing efficiency, $s = 2^j, \quad j = 1, 2, \ldots, J$ (Mallat, 1999). Thus, equation 2.22 gets the form of a discrete convolution (e.g. Katul and Parlange, 1994)

$$Wf^{(j)}[p] = \sum_{p=1}^{2^{j-j}} g^{(j)}[p - 2^j t] f[t],$$ (2.23)

where $j$ is the scale index $[1, J]$, $Wf^{(j)}$ are the the wavelet coefficient at scale $j$, $p$ the position index $[1, 2^{j-j}]$, $g^{(j)}$ is the discrete version of $\psi(t)$ at scale $j$ and $f[t]$ the discrete time series of a variable of interest. The convolution is similar to cross-correlation: it gives the weighted average of $f[t]$ at the moment $p$ where the weightening is given by the wavelet $g[-t]$ that has been shifted by $p$. Note that the wavelet is zero around the desired area. The wavelet coefficients are analogous to the amplitudes in the DFT.

The wavelet transform here has a couple of important features that must be
mentioned. First, the energy is conserved in the wavelet transform, that is

\[
\sum_{t=-\infty}^{\infty} f(t)^2 = \sum_{j=-\infty}^{\infty} \sum_{p=-\infty}^{+\infty} (Wf^{(j)}[p])^2.
\] (2.24)

Second, the discrete wavelet is orthogonal to its dilations and translations

\[
\sum_{p=-\infty}^{\infty} g^{(j)}[p - 2^j t]g^{(l)}[p - 2^l t] = \begin{cases} 1, & \text{if } j = l \text{ and } t = y \\ 0, & \text{otherwise}. \end{cases}
\] (2.25)

This is important for reducing undesired relations between wavelet coefficients. If the wavelets were not orthogonal, correlations between wavelet coefficients of a time series may be observed instead of the correlation of the actual time series (e.g. Katul and Parlange (1994)). The limits of infinity above are of course not reached in a discrete time series.

The resolution of the DWT depends on the data length used in the computations. At high frequencies the wavelet is very narrow and at low frequencies it is broad. Thus, the resolution at lower frequencies is much lower. For instance, if the data length \(N = 2^{14} = 16384\), 14 different spatial scales [m] are resolved (Katul et al., 1994)

\[
R_j = 2^j dy = \frac{2^j \cdot U}{f_s}, \quad j = 1, 2, \ldots, 14
\] (2.26)

Here \(U\) is the mean wind speed and \(f_s\) the sampling frequency. In such a case, all the scales will not have the same resolution. The smallest scale will have \(2^{13}\) wavelet coefficients whereas the coarsest scale only 1 coefficient. The sum of the number of wavelet coefficients in all scales equals the data length for which the DWT was applied. In other words, if DWT is done to a data record of length \(N\), a vector of the same length is returned. A continuous wavelet transfer, on the other hand, would return \(N^2\) wavelet coefficients. Figure 2.3 elucidates the dyadic structure of wavelet coefficients. The five last scale indices \((j=10-14)\) and their corresponding dilation parameters \((s = 2^j)\), spatial scales \((R_j, \text{assuming } U = 4 \text{ m/s and } f_s = 10)\) and position indices \((p = 2^{j-j})\) are shown.

The dissimilarity in scale resolution stems from the calculation procedure based on filtering: the DWT actually decomposes the signal into low-pass and high-pass components subsampled by 2 (Mallat, 1999). The low-pass filtering smoothens the
Figure 2.3: The five largest scales of a data record with a length $2^{14}$. The scale indices ($j$), dilation parameters ($s$), spatial scales ($R_j$, $U = 4$ m/s and $f_s = 10$) and position indices ($p$) are calculated. The vector index shows the position of the values in a wavelet coefficient vector that is returned when a wavelet transform is done.

data whereas the high-pass filter conserves the variation. A new filtering is done to the low-pass filtered data and the filtering goes on until the low-pass filtering gives zero coefficients. Different kind of filters are found in filter banks. The choice of a wavelet is actually a choice of filters as the wavelets appear in the computation in the form of filters. The inverse transform is done by reconstructing the original data vector starting from the coarsest scales. Finer and finer variation is gradually added until the original time series is gained.

As for the DFT, the power spectral and co-spectral densities, just to name but a few, can be derived from the wavelet coefficients. The power spectral density of a certain scale is the total energy in the scale divided by the change in wave number $\Delta k_j = 2\pi \ln(2)/R_j$ (e.g. Katul and Parlange (1994))

$$S^{(j)} = \frac{1}{N\Delta k_j} \sum_{p=1}^{2^{j-j}} (Wf^{(j)}[p])^2 = \frac{2^j dy}{N2\pi \ln(2)} \sum_{p=1}^{2^{j-j}} (Wf^{(j)}[p])^2$$

$$= \frac{dy}{2\pi \ln(2)} \langle (Wf^{(j)}[p])^2 \rangle.$$  

The standard deviation of the power spectral density is

$$\sigma^{(j)} = \frac{dy}{2\pi \ln(2)} \left[ \langle (Wf^{(j)}[p])^4 \rangle - \langle (Wf^{(j)}[p])^2 \rangle^2 \right]^{1/2},$$

where $\langle * \rangle$ denotes now an average over all position indices. The co-spectrum is calculated from the wavelet coefficients of two variables $A$ and $B$ as

$$C^{(j)} = \frac{dy}{2\pi \ln(2)} \langle Wf^{(j)}_A[p] \cdot Wf^{(j)}_B[p] \rangle.$$
2.4 Similarity theory and forms of spectra

Before going into similarity theories, the energy spectrum of turbulence has to be introduced. The energy spectrum of turbulence in the surface layer consists of three distinct regions. In the energy-containing range turbulence is produced thermally and mechanically and the range contributes most to the energy. In the inertial subrange turbulence is neither produced nor destroyed but it is transported down to smaller scales. This is called the cascade process. In the dissipation range turbulent kinetic energy is finally transformed into heat. The dissipation rate $\epsilon$ must equal the cascade rate (Kaimal and Finnigan, 1994).

![Energy spectrum of atmospheric turbulence](chart)

Figure 2.4: Schematic of the energy spectrum of atmospheric turbulence as a function of wave number in the boundary layer. After Kaimal and Finnigan (1994), figure 2.1.

2.4.1 Kolmogorov similarity

In the inertial subrange, $\epsilon$ is the only determining variable and the shape of the energy spectrum can be derived with dimension analysis (Kolmogorov, 1941). Because $u'^2 = \int_{-\infty}^{\infty} E(k_1)dk_1$, $E(k_1)$ must have the dimension of $m^3s^{-2}$, and the energy
can be expressed
\[ E(k_1) = \alpha_u \epsilon^a \cdot k_1^b, \]  
where \( E(k_1) \) is the one-dimensional spectral density and \( k_1 \) is the corresponding wave number. The values of \( a = 2/3 \) and \( b = -5/3 \) can be obtained through dimension analysis and \( \alpha_u = 0.55 \) is the Kolmogorov constant for one-dimensional \( u \) spectra. It is a universal constant and determinable from measurements. The equation implies that a logarithmic plot with \( E(k_1) \) versus \( k_1 \) has a slope of \(-5/3\) in the inertial subrange. Using Taylor’s assumption of frozen turbulence \( (k = \frac{2\pi}{N} = \frac{2\pi n}{U}) \), equation \( 2.31 \) for horizontal wind converts to
\[ k_1 E(k_1) = n S_u(n) = \frac{\alpha_u}{(2\pi)^{2/3}} \epsilon^{2/3} n^{-2/3} u^{2/3}. \]  
Now it can also be seen that a logarithmic plot with \( n S(n) \) versus \( n \) has a slope of \(-2/3\) in the inertial subrange.

In the dissipation range, there are two determining variables, namely \( \epsilon \) and \( \nu \), the viscosity. As a consequence, dimension analysis cannot be used for determining the shape of the energy spectrum.

### 2.4.2 Monin-Obukhov similarity theory

The Monin-Obukhov similarity theory [Monin and Obukhov 1954] describes the vertical behavior of non-dimensionalized mean flow turbulence in the surface layer. The relationships are all functions of the stability parameter \( \zeta = z/L \), which is a function of the key parameters: height above ground \( z \), buoyancy \( g/\theta_v \), surface stress \( \tau \) and heat flux at the ground \( H \), as shown in section 2.2.

The Obukhov length can be used for determining the dimensionless gradients of wind speed, potential temperature and humidity
\[ \frac{d\overline{u} \kappa z}{dz} \frac{z}{u_*} = \phi_m \left( \frac{z}{L} \right), \quad \frac{d\overline{\theta} \kappa z}{dz} \frac{z}{T_*} = \phi_h \left( \frac{z}{L} \right), \quad \frac{d\overline{q} \kappa z}{dz} \frac{z}{q_*} = \phi_q \left( \frac{z}{L} \right). \]  
Here \( \kappa_v = 0.4 \) is the von Kármán constant. The functions \( \phi \) can be determined experimentally and are universal, that is, they do not depend on the measurement time or location. In neutral conditions, all the functions \( \phi \) equal unity and a logarithmic law for wind speed can be derived
\[ \overline{u}(z) = \frac{u_*}{\kappa_v} \ln \left( \frac{z}{z_0} \right). \]  

---

21
Here $z_0$ is the roughness length and it will be further explained in section 2.5.1.

Also the respective power spectra for $u$, $\theta$ and $q$ can be noted as functions of $\zeta$ and the dimensionless frequency $f = n z / U$. Combining the knowledge from Kolmogorov and the MO-theory leads to

\[
\frac{nS_u(n)}{u_*^2} = \frac{\alpha_u}{(2\pi \kappa_v)^{2/3}} \left( \frac{\kappa_v \frac{\zeta}{u_*^3}}{w_z^3} \right)^{2/3} \left( \frac{n z}{U} \right)^{-2/3} = \frac{\alpha_u}{(2\pi \kappa_v)^{2/3}} \phi_{\zeta}^{2/3} f^{-2/3} \tag{2.35}
\]

\[
\frac{nS_u(n)}{u_*^2 \phi_{\zeta}^{2/3}} = 0.3 f^{-2/3}, \tag{2.36}
\]

where

\[
\phi_{\zeta}^{2/3} = \begin{cases} 
1 + 0.5|\zeta|^{2/3} & \zeta \leq 0 \\
(1 + 5|\zeta|)^{2/3} & \zeta \geq 0.
\end{cases} \tag{2.37}
\]

This means that all power spectra independent of the stability fall together in the inertial subrange. Similarly for other wind components and heat

\[
\frac{nS_{v,w}(n)}{u_*^2 \phi_{\zeta}^{2/3}} = \frac{4}{3} \frac{\alpha_u}{(2\pi \kappa_v)^{2/3}} f^{-2/3} = 0.4 f^{-2/3} \tag{2.38}
\]

\[
\frac{nS_\theta(n)}{T_*^2 \phi_{\zeta}^{-1/3} \phi_h} = \frac{\beta_1}{(2\pi \kappa_v)^{2/3}} f^{-2/3} = 0.43 f^{-2/3} \tag{2.39}
\]

where $\beta_1 = 0.8$ is the Kolmogorov constant for one-dimensional temperature spectra (similar to $\alpha_u$) and

\[
\phi_h = \begin{cases} 
(1 + 16|\zeta|)^{-1/2} & \zeta \leq 0 \\
(1 + 5\zeta) & \zeta \geq 0.
\end{cases} \tag{2.40}
\]

Here it has been assumed that $\phi_N = \phi_h$, that is, there is a local balance between the production and destruction of temperature variance. From above one can see the local isotropy, $S_v(n) = S_w(n) = (4/3)S_u(n)$, that is, the velocity field is independent of rotation and reflection about its spatial axis in the inertial subrange (Kaimal and Finnigan, 1994).

Figure 2.5 shows these four power spectra for different stabilities. The vertical wind obviously exhibits most systematic behavior related to $\zeta$. The spectral peak is shifted towards smaller frequencies as stability decreases. In other words, larger turbulence is responsible for turbulent transfer in unstable conditions compared with stable conditions. In the unstable range, the spectral peak stops shifting to lower frequencies and the spectra becomes insensitive to $\zeta$. The other spectra ($u, v, \theta$)
progress systematically in the stable range whereas the unstable range shows more
spread. Consequently, the unstable u and v spectra do not follow the MO-similarity.

Also the co-spectra have predicted shapes according to MO-similarity

\[-\frac{nC_{wu}(n)}{u_0^2G(\zeta)} = 0.05f^{-4/3}
\]

\[-\frac{nC_{w\theta}(n)}{u_0T_\star H(\zeta)} = 0.14f^{-4/3}
\]

\[G(\zeta) = \begin{cases} 1 & -2 \leq \zeta \leq 0 \\ 1 + 7.9\zeta & 0 \leq \zeta \leq 2 \end{cases}
\]

\[H(\zeta) = \begin{cases} 1 & -2 \leq \zeta \leq 0 \\ 1 + 6.4\zeta & 0 \leq \zeta \leq 2 \end{cases}
\]

These co-spectra are represented in figure 2.6. In the stable range the co-spectra
separate according to \(\zeta\) as did the power spectra. In contrast, in the unstable
range \((-2 \leq \zeta \leq 0)\) the co-spectra are found in a narrow band that straddles
the neutral co-spectra. Consequently one form can be used for all co-spectra for
unstable stratification (see section 2.4.3). It can also be seen from figure 2.6 that
the co-spectra follow a -4/3 slope in the inertial subrange. The momentum co-
spectra attains the slope at lower frequencies and thus it can be stated that heat is
transported in smaller eddies (higher frequency) than momentum.

The Monin-Obukhov similarity theory does not apply when the atmosphere is
highly convective. In such a case \(u_0\) is no longer an important scaling variable. The
theory does not either apply in the roughness sublayer found above a canopy (section
2.5.2). This shortcoming of the theory was first described by Raupach et al. (1980).
The theory also requires horizontal homogeneity and that \(|\zeta|\leq 1 - 2\). Even under
ideal conditions, the accuracy of the theory is only about 10-20% (Foken, 2006).

2.4.3 Forms of spectra

Approximative forms of the co-spectra represented earlier are often used. The nor-
malized co-spectra for scalars for an unstable surface layer (strictly speaking the
Figure 2.5: Normalized power spectra of three wind components and potential temperature for different degrees of stability ($\zeta = z/L$). Note that here $n$ is dimensionless frequency and $f$ natural frequency. After Kaimal and Finnigan (1994).

Neutral limit of the stably stratified surface layer are (Horst, 1997)

$$\frac{nC_{ws}(f)}{w's^2} = \begin{cases} 
1.05 f/f_m (1 + 1.33 f/f_m)^{7/4}, & f \leq 1.0 \\
0.387 f/f_m (1 + 0.38 f/f_m)^{7/3}, & f \geq 1.0 
\end{cases}$$

(2.45)

and for stable stratification

$$\frac{nC_{ws}(f)}{w's^2} = \frac{0.637 f/f_m}{(1 + 0.91 f/f_m)^{2.1}}.$$  

(2.47)

Here the normalized frequency that has the maximum contribution to the flux is denoted by $f_m = n_m z/U$. $f_m$ is a constant for unstable situations, but for stable situations it is a function of stability according to Rannik et al. (2004)

$$f_m = a_1 (1 + a_2 \zeta^{a_3}).$$

(2.48)
Figure 2.6: Normalized co-spectra (equation 2.44) of momentum ($u'w'$) and sensible heat ($w'\theta'$) for different degrees of stability ($\zeta = z/L$). Note that here $n$ is dimensionless frequency and $f$ natural frequency. After Kaimal and Finnigan (1994).

Similar forms for power spectra can also be defined. The normalized power spectra for a scalar during unstable stratification is (Rannik and Vesala, 1999)

$$\frac{nS_s(f)}{\sigma_s^2} = \begin{cases} 
0.934 f/f_m & f < 0.15 \\
0.427 f/f_m & f \geq 0.15
\end{cases}$$

(2.49)

and for stable stratification

$$\frac{nS_s(f)}{\sigma_s^2} = \frac{0.644 f/f_m}{(1 + 1.5 f/f_m)^{5/3}}.$$  

(2.50)

$f_m$ is again a constant for unstable situations and depends on stability according to 2.48 in stable situations. The stability dependency is also evident in the above figure 2.5.

Note that the power and co-spectra here are normalized by the covariance or the standard deviation as an opposite to the spectra in equations 2.39 and 2.44. Consequently, the integrals of the approximative forms here are one whereas the integrals of the spectra in figures 2.6 and 2.5 are not.
2.4.4 Spectral corrections

The fluxes measured with the EC system tend to be underestimations of the unattenuated fluxes in spite of careful measurement design. Flux is lost in the low frequency domain due to running mean or linear de-trending (Rannik and Vesala, 1999) and a limited averaging period (Sakai et al., 2001 and Finnigan et al., 2003). The co-variance losses at high frequencies stem from the incapability of a measurement system to resolve variations caused by small eddies. In the transfer function method, the relationship between the measured and the real covariance can be expressed with a frequency dependent transfer function $TF(n)$ as follows

$$\frac{w's'_{\text{meas}}}{w's'} = \int_0^\infty TF(n) \cdot C_{ws}(n)dn \int_0^\infty C_{ws}(n)dn = CF,$$

(2.52)

where $w's'_{\text{meas}}$ is the measured covariance, $w's'$ the unattenuated covariance and $CF$ is the correction factor. The method was originally initiated by Moore (1986) and further developed by Horst (1997), Horst (2000), Massman (2000) and Massman (2001). The transfer function consists of a low ($TF_{LF}$) and a high ($TF_{HF} = \prod TF_{HF,i}$) frequency part: $TF_{LF}$ is zero at low frequencies and rises to one at high frequencies whereas $TF_{HF}$ behaves in an opposite way. $TF_{HF}$ for a closed-path IRGA can also be represented in the form (Horst, 1997)

$$TF_{HF}(n) = \frac{1}{1 + (2\pi n \tau_s)^2},$$

(2.53)

where $\tau_s$ [s] is the response time of the system which is equivalent to the inductance in an alternating circuit (Eugster and Senn, 1995). The cut-off frequency is the frequency where $TF_{HF}$ has decreased by a factor of two (Aubinet et al., 2000). According to recent studies, the response time is a function of relative humidity (Ibrom et al., 2007, Massman and Ibrom, 2008 and Mammarella et al., 2009) and depends on tube age (Leuning and Judd, 1996, Su et al., 2004 and Mammarella et al., 2009). Also Clement (2004) showed the exponential dependency of lag time on relative humidity but did not combine this information with high frequency attenuation and $\tau_s$.

From equation 2.53 one can see that there is an additional $f^{-2}$ dependency for dampened co-spectra and power spectra. This leads to a slope of $-10/3$ instead of the traditional $-4/3$ for the co-spectra and $-8/3$ instead of $-2/3$ for power spectra.
in the inertial subrange. The transfer function can be determined both theoretically and experimentally. The frequency corrections in this work will be concentrated on the experimental correction of closed-path infra-red gas analyzer data (section 3.4).

2.5 Urban boundary layer

2.5.1 Effects of buildings

The inclusion of buildings in a surrounding affects the radiative, thermal, moisture and aerodynamic characteristics of the area. The most prominent effect on radiation is the reduction of received global radiation by areas in shadow but then again there is an increase in global radiation receipt due to reflection from adjacent walls. Net long wave cooling is reduced as the downward directed long-wave radiation is increased as a result of warmer surroundings. The upward directed long wave radiation is also decreased, because of a reduced sky factor. The moisture characteristics are mainly altered by the impervious surfaces such as paved roads, parking lots and buildings. This has a profound effect on the energy balance of an urban area (section 2.5.3).

The aerodynamic effects of buildings are numerous. Generally speaking buildings increase the roughness and provide wind shelter. Figure 2.7 is a schematic representation of an airflow encountering a single building. The flow consists of four regimes: undisturbed, displacement, cavity and wake (Oke, 1987). On the lee side of an obstacle, the cavity flow consists of two horizontal circulations (not shown) and in the wake flow the residual turbulence facilitates a greater momentum transport to the ground. Figure 2.8 shows the behavior of flow in a many-building system. Three different flow patterns can be distinguished depending upon the height to width ratio (H/W), where H is the mean building height and W the distance between buildings. When the buildings are sparsely spaced (H/W<0.4) the flow patterns appear the same as for isolated buildings and the flow is called the isolated roughness flow. When the buildings are closer to each other (0.4<H/W<0.7) the wake of the first building interferes with the flow pattern of the next building and a wake interference flow develops. Finally, when the spacing is further decreased the flow skims over the buildings and a lee vortex develops between the adjacent buildings resulting in a skimming flow. A new surface at the height of \( z_r \) is developed. Skimming flow
can only develop when the buildings have approximately the same height \cite{Hanna2002}. Generally roughness increases with decreasing building spacing, but \cite{Hanna2002} showed that at a certain point the increasing stops and the roughness actually decreases. Air can also be stagnated between buildings which results in bad air quality \cite{Oke1987}.

Figure 2.7: Schematic of a flow encountering a building. The flow zones are A – undisturbed, B – displacement, C – cavity and D – wake. After \cite{Oke1987}, figure 8.1.

Figure 2.8: Flow patterns associated with different urban geometries, after \cite{Oke1987}, figure 8.2.

The logarithmic wind profile introduced earlier (equation 2.34) is altered by the roughness described above. Figure 2.9 shows the wind profile over a flat surface (left) and over an urban surface (right). The latter profile deviates from the logarithmic law roughly below a couple of meters above the roof tops. The wind speed according to the logarithmic profile would go to zero at the height \(z_d + z_0\), where \(z_d\) is the zero displacement height and \(z_0\) the roughness length. Consequently, atmospheric stability has to be redefined as \(\zeta = \frac{z - z_d}{L}\) for rough surfaces. The roughness length is a measure of the mechanical mixing caused by the roughness elements and is
approximately equal to $0.1 \cdot z_H$ (Hanna and Britter, 2002). It also depends on the shape and orientation of the roughness elements. The logarithmic wind profile is valid in neutral conditions in densely built areas when $z > 1.5 z_H$ and $z$ is less than the boundary layer height (Hanna and Britter, 2002). The roughness sublayer height, $z_r$, in figure 2.9 will be explained later.

Stability is also affected by the increased roughness at urban sites. The Monin-Obukhov length is increased due to the strengthened $u_*$ as $L(u_*^3)$. This leads to more neutral conditions as stability $\zeta L^{-1}$. Anthropogenic heating contributes also to the stability and the subject will be treated in section 2.5.4. The altering of $L$ gives rise to doubt of the applicability of the turbulence similarity theories at an urban site. Surprisingly, it has been proven that urban turbulence can be interpreted in the same framework as turbulence above more homogeneous surfaces (Roth, 2000). Anyhow, wake turbulence must be taken into account in sparse canopies (Roth, 2000).

Figure 2.9: Schematic of a logarithmic wind profile at a flat surface (left) and at an urban site (right). $z_0$ – roughness length, $z_d$ – displacement height, $z_r$ – roughness sublayer height, $z_H$ – average height of roughness elements (trees, buildings).

### 2.5.2 Vertical scales

The boundary layer associated with an urban area has several vertical scales and concepts that must be understood before further going into studying the phenomena within the boundary layer. Many of the following vertical scales are subjects of much debate.

When air flows from the countryside to the city, an *internal boundary layer* develops downwind from the leading edge of the city. This surface separates the planetary
boundary layer from the urban boundary layer (UBL) and the urban canopy layer (UCL) (figure 2.10). The boundary layer below the internal boundary layer, here UBL, is adjusted to the new surface and grows from bottom up. Oke (1976) was the first to introduce a division to UBL and UCL. According to Oke (1987), 'the UBL is a local to meso-scale phenomenon whose characteristics are governed by the nature of the general urban surface' whereas the UCL is 'produced by micro-scale processes operating in the streets between buildings'.Arnfield (2003) further emphasizes that the division reflects different assemblages of processes rather than mere scales.

The UBL can further be divided into a mixing layer and a surface layer. In the mixing layer, turbulence is vigorous and the air is almost uniformly mixed in the vertical direction. Such properties as potential temperature and wind speed are constants with height (figure 2.12, e.g. Stull (1988)). In the surface layer, air is not uniformly mixed and mechanical generation of turbulence exceeds buoyant generation or consumption..

The surface layer consists of a roughness sublayer (RSL, Raupach (1979), a.k.a. turbulent wake layer, transition layer, interfacial layer) and an inertial sublayer (ISL) (figure 2.10). In the RSL, which includes the UCL, the flow consists of wakes and plumes caused by individual roughness elements and the MO-theory is invalid. In addition, the air volume is characterized by strong vertical shear, large turbulent intensities and local advection and the layer might extend up to several times the average building height (figure 2.9, Roth (2000), an overview). A rule of thumb is that \( z_r \) extends from \( 1.5 z_H \) in densely built areas to over \( 4z_H \) in sparsely built areas (Grimmond and Oke, 1999a). The inertial sublayer is characterized by constant fluxes with height and standard turbulence theories are applicable. The flow structure and thermodynamic properties are adapted to the surface type and thus micro-meteorological measurements are representative of the underlying surface. To the contrary, measurements below \( z_r \) register microclimate anomalies (Oke, 2006).

### 2.5.3 Urban energy balance

A fundamental problem arises when trying to evaluate the urban energy balance as distinct to other surfaces, namely, the definition of a surface itself. Oke (1988) introduced a new way to treat the problem. They replaced the traditional concept of a massless surface with that of a building-air volume (figure 2.11). The top of
Figure 2.10: The structure of the urban boundary layer. PBL – planetary boundary layer, UBL – urban boundary layer, UCL – urban canopy layer, RBL – rural boundary layer, RSL – roughness sublayer, ISL – internal sublayer, after Oke (1982). Note that the proportions in the figure are not in scale.

The volume is at the height of \( z_0 + z_d \), which corresponds to a level slightly above roof tops, whereas the bottom is at a depth below surface where heat flux goes to zero. The energy balance can now be defined as (Oke (1987); Offerle et al. (2005))

\[
R_n + Q_F = H + LE + Q_S + Q_A + \epsilon_{EB},
\]

(2.54)

where

- \( R_n \) net all-wave radiation
- \( Q_F \) anthropogenic heat release
- \( H \) sensible heat flux
- \( LE \) latent heat flux
- \( Q_S \) change in heat storage (includes air, trees, buildings, soil etc.)
- \( Q_A \) advection trough the volume
- \( \epsilon_{EB} \) error term, includes all other processes

Net radiation can further be decomposed to

\[
R_n = SW \downarrow - SW \uparrow + LW \downarrow - LW \uparrow = (1 - \alpha)SW \downarrow + LW \downarrow - \varepsilon \sigma T^4,
\]

(2.55)
where $SW$ denotes short-wave and $LW$ long-wave radiation. The arrows indicate the directions of radiation propagation and $\alpha$ is the albedo of the surface, $\varepsilon$ the emissivity and $T$ the temperature in Kelvins. In equation 2.54 the advection term is often neglected as it is hard to measure though advection is more a norm than an exception in an urban environment (Roth, 2000). Advection is especially caused by the horizontal temperature gradient within a city (section 2.5.4). The error term, including all other processes, is not taken into account since it is hard to determine. An example of an excluded process could be rain water which flows out of the city via sewers and takes energy with it. The equation differs from the traditional energy balance equations in the inclusion of $Q_F$. The term includes such heat sources as vehicles, houses and metabolism. The term is directly controlled by human activities and indirectly by nature. For example, the need of heating follows an annual cycle, which stems from the input of solar radiation.

When advection is neglected, $Q_F$ and $Q_S$ can be combined to form a residual term

$$Res = R_n - H - LE = Q_S - Q_F.$$  

(2.56)

The residual term includes now implicitly the error term. This variable is negative when there is an additional energy source fueling $H$ and $LE$. The source is either a heat release from the building-air volume ($Q_S < 0$) or an anthropogenic heat flux ($Q_F > 0$) or both. Conversely, $Res$ can also be positive when more heat is stored to the building-air volume than is release from anthropogenic activities ($Q_S > Q_F$). The neglection of $\varepsilon$ must be kept in mind since it is well known that the traditional energy balance (without $Q_F$) at more simple cites as forests is not closed. Wilson et al. (2002), for instance, found an average 20% lack of closure at 22 FLUXNET sites. This means that $Res$ is probably always too large.

The terms in equation 2.54 for an urban site differ generally from those for a rural site. In addition to the factors caused by buildings (section 2.5.1), net radiation is altered by pollutants as they decrease $SW \downarrow$ and increase $LW \downarrow$. Furthermore, the albedo, and therefore $SW \uparrow$, is decreased as buildings have typically dark surfaces in high latitude cities. Vertical walls that are snow-free during winter also substantially decrease the albedo. $LW \uparrow$ is increased owing to higher surface temperature. For these reasons, the net effect of altering $R_n$ in an urban environment is small (e.g. Rotach et al. (2005); Christen and Vogt (2004)). The latent heat flux is highly restricted by the availability of water in an urban environment. The patchy occur-
Figure 2.11: A schematic describing the concept of building-air volume and the fluxes in the energy balance (Equation [2.54]). UCL – urban canopy layer, $H$ – sensible heat flux, $LE$ – latent heat flux, $Q_S$ – heat storage to the building-air volume, $R_n$ – net radiation, $Q_F$ anthropogenic heat flux, $Q_A$ – advection. After Oke (1988).

The occurrence of vegetated surfaces (transpiration) and wet surfaces (evaporation) makes the distribution of sources and sinks very heterogeneous. The imperviousness of some surfaces also affects $LE$. The magnitude of $Q_S$ is more significant to the urban energy balance than is for that of other land covers (Grimmond and Oke, 1999b). Consequently, $Q_S$ causes inertia to the system and dampens the variability of turbulent fluxes. In addition, $Q_S$ together with an increased roughness in cities leads to a more neutral atmospheric stability over cities during night.

### 2.5.4 Climate of the UBL

The difference between energy partitioning at urban and rural sites results in a modified climate in urban areas. Essentially, the difference arises from the altered thermal, aerodynamic, radiative and moisture characteristics of an urban area.

The heat island effect is probably the most well documented example of inadvertent climate modification (Arnfield, 2003). The phenomena is characterized by higher long-term temperatures in a city as an opposite to the surrounding rural ar-
eas. In a diurnal sense, nocturnal temperatures are higher and day-time values can actually be lower in a city. The heat island intensity is defined as the temperature difference between the urban and rural sites \( T_{urban} - T_{rural} \) (Oke (1987)) and the intensity can extend up to 10°C degrees Celsius during winter nights (e.g. Grimmond (2007)). Recent research shows that the main reason for the temperature difference is the influence of building geometry on radiative fluxes (Souch and Grimmond (2006), an overview). Other causes are the anthropogenic heat release from the city and an increased sensible heat flux input. The suppressed LE in cities causes the energy input to be directed to \( H \) which leads to an increased input of \( H \) from the canopy. Sensible heat is also entrained from above as a result of increased turbulence due to larger roughness (Oke, 1982). For instance, warm air can be entrained from an overlying inversion during night time. The anthropogenic heat release plays an important role especially in high latitude settlements during winter.

The heat island intensity depends on several factors. Oke (1987) already suggested that the intensity is inversely proportional to wind speed and cloudiness and the statements are still valid (Arnfield, 2003). The intensity is greater during anticyclonic conditions (Oke, 1982) and the heat island is least developed in summer (Souch and Grimmond, 2006). Oke (1987) also suggested that the intensity is proportional to the logarithm of the city population, but more recent work has shown that the dependence stems essentially from building geometry and the used building fabric (Klysik and Fortuniak, 1999).

The heat island effect leads to an altered temperature profile both within a city and downwind of the city. Figure 2.12 a) is a schematic of potential temperature profiles in rural and urban areas during night. A rural inversion exists on the upwind side of the urban area. The urban area, on the other hand, is characterized by a weakly convective layer close to the ground. The nocturnal urban boundary layer may be 500 m high depending on stability and roughness characteristics. The effect of a city on a rural area can be seen as a heat plume extending several kilometers downwind of a it. During day-time, the rural profile is unstable close to the ground, neutral in the mixed layer and has an inversion showing the top of the boundary layer (figure 2.12 b). The corresponding urban profile is more unstable up to about 250 m and a higher mixed layer lies on top of it. The UBL is generally higher than the rural boundary layer due to the heat island effect and can extend up to 1.5 km during day-time. A heat plume can also be seen during day-time. (Oke 1982)
Figure 2.12: a) The height of the urban boundary layer and the rural boundary layer during night in addition to potential temperature profiles. b) Same for daytime, after Oke (1982). (figure 2 a and b).

An urban area also has an effect on day-to-day weather. The temperature gradient caused by the heat island effect together with the roughness change between rural and urban environments can cause an urban circulation pattern where air is converged in the city leading to an up-draft (Oke, 1987). According to Shephard (2005, an overview), additional rain-producing clouds can develop in urban areas due to the increase of hygroscopic nuclei, enhanced turbulence and convergent airflow over the area. In addition, the addition of water vapor from anthropogenic sources may enhance precipitation. Several attempts have been done to take urban areas appropriately into account in numerical weather prediction models. Best (2005) for
instance introduced a new scheme that simulates the near-neutral nocturnal boundary layer, but does still not take anthropogenic heat release into consideration.

2.5.5 Spectra in an urban environment

Roth (2000) made a comprehensive analysis of urban spectra compared to their rural counterparts. The analysis was made for velocity components and temperature in neutral and unstable atmospheric conditions. Stable stratification was not included due to the lack of data since urban environments are rarely stably stratified. The velocity power spectra are generally in good agreement with the widely accepted models derived from data measured at simpler terrains. Only close to and within an urban canopy the spectra are shifted towards higher frequencies and a clear maximum is lost. The peak of vertical velocity power spectra is observed to shift to lower frequencies at all heights indicating a turbulence production by wakes. Temperature power spectra in unstable situations show wide scatter but no systematic behavior was observed. To the contrary, the temperature co-spectra collapse together and agree well with the reference data in spectral shape and peak position. Momentum co-spectra, on the other hand, are ill-defined with a high degree of scatter.
Materials and methods

In this section, the features of the measurement site and measurement system itself will be described. Furthermore, the EC data processing will be gone through briefly. More emphasis will be set on the description of the high frequency spectral corrections. The traditional transfer function method will be explained and a new correction method based on wavelet analysis will be introduced.

3.1 Site description

The urban SMEAR III station (Station for Measuring Ecosystem-Atmosphere Relationships) is a member of the series of SMEAR stations in Finland, established in the fall of 2004. The station consists of two units in the Helsinki city: ecosystem research is done at the Viikki campus 7 km north-east of the city center and micrometeorological turbulence and air quality measurements are done at the Kumpula campus 4 km north-east from the center (60°12’N, 24°58’E). This study concentrates on data from the latter location.

Helsinki is located at the coast of the Gulf of Finland. The weather varies from continental to marine depending upon the prevailing wind direction. The 30 year monthly mean temperature ranges from -4.9°C in February to 17.2°C in June and the annual average is 5.6 °C. The average yearly precipitation is 642 mm (Drebs et al., 2002). The Helsinki metropolitan area has ca. 1 million inhabitants and is 762 km² in size.

The eddy covariance measurements at the Kumpula campus are done from a 31 m high tower 26 m a.s.l. whereas the meteorological data acquisition is done
on the top of a university building (figure 3.1). The area around the tower is highly heterogeneous consisting of buildings, roads, parking lots, trees, shrubs and grass. According to Vesala et al. (2008), the area can be divided into three sectors: urban (320°-40°), road (40°-180°) and vegetation (180°-340°), see figure 3.1. The anemometer is in the wake of the tower when the airflow is from 0-60°. The average building height in the urban sector is 20±2m and thus the height of the roughness sublayer (z_r) is at least 30 m which means that the turbulence measurements from the 31 m tower might represent microclimate anomalies (section 2.5.2). The near by road is fairly busy with its mean rate of 50000 cars per day (Järvi et al., 2009a). The land usage fractions of each sector are reported in table 3.1. Also included is the displacement height per sector defined as 2/3 of the main canopy height (Grimmond and Oke, 1999a).

Flux footprint area modeling of the road and vegetation sectors was done by

Figure 3.1: Footprint modeling results of the area around the EC measurement tower (black star). Black contours show the topography, white contours give the footprint for a) canopy and b) soil sources. Scale 10^{-4}m^{-2}. Red lines denote boundaries between the three wind direction sectors. The white star denotes the university building from the top of which the meteorological measurements are done. After Vesala et al. (2008)
Table 3.1: The sector types, their wind directions, displacement heights \( (z_d) \) and land usage fractions. The values are for an area within 250m from the tower. Also shown are the fractional contributions of building (B), vegetation (V) and road (R) area to the footprint of the canopy and the soil. Only results for the road and the vegetation sector are shown. After [Vesala et al. (2008)].

<table>
<thead>
<tr>
<th>Sector</th>
<th>Sector type</th>
<th>( z_d )</th>
<th>Fraction</th>
<th>Footprint</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>B</td>
<td>V</td>
</tr>
<tr>
<td>320°-40°</td>
<td>Urban</td>
<td>13 m</td>
<td>0.42</td>
<td>0.51</td>
</tr>
<tr>
<td>40°-180°</td>
<td>Road</td>
<td>8 m</td>
<td>0.10</td>
<td>0.60</td>
</tr>
<tr>
<td>180°-320°</td>
<td>Vegetation</td>
<td>6 m</td>
<td>0.02</td>
<td>0.13</td>
</tr>
<tr>
<td>All</td>
<td></td>
<td></td>
<td>0.14</td>
<td>0.40</td>
</tr>
</tbody>
</table>

Vesala et al. (2008). They used a one-and-a-half-order turbulence closure model SCADIS (Sogachev et al. (2004), Sogachev and Lloyd (2004), Sogachev and Panferov (2006)), which assumes neutral stratification and does not take into account anthropogenic heating. The footprints were modeled separately for sinks and sources located on the ground and in the canopy. Due to the high heterogeneity, the footprint appeared asymmetric along the chosen wind directions (figure 3.1). It can also be noted that the source area of the canopy is much smaller than that of the surface. The chosen flux contributions were 64% and 69% for the road and vegetation sectors. The contribution of three area types (building, vegetation, road) within the sectors are shown in table 3.1. Vegetation has the largest contribution always in the vegetation sector and in the canopy footprint of the road sector. Road has the highest contribution to the soil footprint of the road sector. These results are extremely essential for the analysis of the energy fluxes as a function of wind direction (section 4.3).
3.2 Measurements and data selection

The eddy covariance measurements consisted of an ultrasonic anemometer (USA-1, Metek GmbH, Germany) for measuring the three wind components and the sonic temperature and an open and a closed-path infra-red gas analyzer for measuring water vapor and carbon dioxide (LI7500 and LI-7000, Li-Cor, Inc., Lincoln, Nebraska, USA). All of these three instruments worked with a 10 Hz frequency. In addition, meteorological variables were measured at the top of the university building and the measurements are listed in table 3.2. The four components of radiation were measured also from the EC tower. The source areas of the downward directed sensors represent an area consisting of rock, grass, bushes and trees. Thus, the surface temperature and albedo differ from those of the surfaces in the footprint area of the fluxes. This must be kept in mind when calculating the residual of the energy balance.

Table 3.2: List of measurements of meteorological variables at the SMEAR III station.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Sensor</th>
<th>Equipment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air temperature</td>
<td>platinum resistant thermometer</td>
<td>Pt-100</td>
</tr>
<tr>
<td>Net radiation</td>
<td>net radiometer</td>
<td>CNR1, Kipp and Zonen, Delft, Netherlands</td>
</tr>
<tr>
<td>Air pressure</td>
<td>barometer</td>
<td>Vaisala DPA500, Vaisla Oyj, Vantaa, Finland</td>
</tr>
<tr>
<td>Relative humidity</td>
<td>platinum resistance thermometer and thin film polymer sensor</td>
<td>Vaisala HMP243, Vaisala Oyj, Vantaa, Finland</td>
</tr>
</tbody>
</table>

The EC measurements with the open-path IRGA started already 1, December 2005 whereas the closed-path IRGA was installed 25, June 2007. Data until the end of June 2009 are used here. The meteorological measurements covered the whole period. There are gaps in the data due to instrument maintenance and measurement malfunctions. For instance, September 2007 for the closed-path IRGA is missing. The years are divided into four seasons according to the daily mean temperature ($T_{ave}$): winter is defined as the period when $T_{ave} \leq 0^\circ C$, summer when $T_{ave} \geq 10^\circ C$ and spring and fall are for $0 < T_{ave} < 10^\circ C$. 
The EC data quality was assessed using the stationarity test introduced by Foken and Wichura (1996). The covariances for the fluxes are calculated for 30-min intervals and as an average of 5-min subintervals. If these two values differ more than 30% from each other, the flux is considered non-stationary, which is against the assumptions in the Monin-Obukhov theory (section 2.4). The rejection percentages due to non-stationarity are in table 3.3. The omittance ranges from 15% for $H$ to 47% for closed-path $LE$. It must be noted, that the open-path data was visually inspected for spikes caused by rain before these quality calculations. The omittance is higher for the energy fluxes during night and the opposite for the momentum flux.

Table 3.3: EC data omittance in percentages with a stationarity limit of 0.30. $\tau$ – momentum flux, $H$ – sensible heat flux, $LE_{OP}$ – latent heat flux with an open-path IRGA, $LE_{CP}$ – latent heat flux with a closed-path IRGA.

<table>
<thead>
<tr>
<th></th>
<th>$\tau$</th>
<th>$H$</th>
<th>$LE_{OP}$</th>
<th>$LE_{CP}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>All</td>
<td>20.89</td>
<td>14.99</td>
<td>25.95</td>
<td>45.34</td>
</tr>
<tr>
<td>Night</td>
<td>18.11</td>
<td>17.79</td>
<td>31.48</td>
<td>57.32</td>
</tr>
<tr>
<td>Day</td>
<td>23.40</td>
<td>12.32</td>
<td>21.59</td>
<td>33.85</td>
</tr>
</tbody>
</table>

3.3 EC data processing

The eddy covariance fluxes were calculated from covariances between a scalar and a vector (equation 2.2) according to widely accepted procedures (Aubinet et al., 2000). First, data is de-spiked and a 2-dimensional coordinate rotation is done (see Vesala et al. (2008) for a comparison between rotations). Second, the time lag between sensors is accounted for by the maximum cross-covariance method. If a maximum is not found, as is the case for weak fluxes, a mean value is used. Third, the data are linearly de-trended. Fourth, the water vapour and CO$_2$ fluxes are corrected for water vapour and heating effects (Webb et al., 1980). Finally, spectral corrections are applied to the data (See section 3.4). The correction according to Burba et al. (2008) was not applied to the open-path water vapour measurements. See Järvi et al. (2009b) for the correction results when applied to CO$_2$.

In the spectral analysis, a Fast Fourier transform (FFT) algorithm is used for making the discrete Fourier transform (section 2.3.1). Before using the algorithm,
the data is linearly de-trended to avoid red noise and a Hamming window is applied for smoothing the data window edges and thus for avoiding leakage (Stull 1988). After calculating the spectra as shown earlier, the data is binned into 75 logarithmically evenly spaced classes to minimize scatter.

3.4 Determining the transfer function

The high frequency transfer function will be defined theoretically for the open-path IRGA data and both theoretically and experimentally for the closed path IRGA data. The low frequency transfer function can only be determined theoretically. The theoretical approaches are represented first as they also give insight to the factors contributing to the variance deficit.

3.4.1 Theoretical transfer functions

Moncrieff et al. (1997) presents the high frequency transfer function as a product of theoretically determinable terms

\[ TF_{HF,\text{wu}} = TF_{d(S)}(n)^2 \cdot TF_w(f_p) \] (3.1)

\[ TF_{HF,\text{w}\vartheta} = TF_{d(S)}(n)^2 \cdot \sqrt{TF_d(n)} \cdot \sqrt{TF_w(f_p)} \] (3.2)

\[ TF_{HF,\text{ws,OP}} = TF_{d(OP)}(n) \cdot TF_{d(S)}(n) \cdot TF_m(n) \cdot TF_w(f_p) \cdot TF_s(f_s) \] (3.3)

\[ TF_{HF,\text{ws,CP}} = TF_{d(CP)}(n) \cdot TF_{d(S)}(n) \cdot TF_m(n) \cdot TF_w(f_p) \cdot TF_s(f_s) \cdot TF_a(n), \] (3.5)

where \( s \) refers to \( \text{H}_2\text{O} \) or \( \text{CO}_2 \) and \( f_s \) and \( f_p \) are described in table 3.4.1. The mathematical definition and the descriptions for the components are as follows (Moore
(1986), Moncrieff et al. (1997), Aubinet et al. (2000))

$$TF_{d(CP)}(n) = \frac{1}{\sqrt{1 + (2\pi n \tau_{CP})^2}}$$
dynamic sensor frequency response, closed-path

$$TF_{d(OP)}(n) = \frac{1}{\sqrt{1 + (2\pi n \tau_{OP})^2}}$$
dynamic sensor frequency response, open-path

$$TF_{d(S)}(n) = \frac{1}{\sqrt{1 + (2\pi n \tau_{S})^2}}$$
dynamic sensor frequency response, sonic

$$TF_m(n) = \frac{1 + (2\pi n)^2 \tau_1 \tau_2}{\sqrt{[1 + (2\pi n \tau_1)^2][1 + (2\pi n \tau_2)^2]}}$$
sensor response mismatch

$$TF_w(f_p) = \frac{2}{\pi f_p} \left[ 1 + \frac{\exp(-2\pi f_p)}{2} - \frac{3[1 - \exp(-2\pi f_p)]}{4f_p} \right]$$
sonic anemometer path averaging

$$TF_t(n) = \frac{1}{2\pi f_p} \left( 3 + \exp(-2\pi f_p) - \frac{4(1 - \exp(-2\pi f_p))}{2\pi f_p} \right)$$
temperature path averaging

$$TF_s(f_s) = \exp(-9.9 f_s^{1.5})$$
sensor separation loss

$$TF_a(n) = \exp\left(-80(Re)^{-1/8}(d/2)n^2A\right)$$
frequency attenuation in the tube

The low frequency transfer function depends on the type of averaging and de-
trending, and it can only be determined theoretically. $TF_{LF}$ for linear de-
trending is defined as (Rannik and Vesala, 1999)

$$TF_{LF} = 1 - \left( \sin(\pi n T) \over \pi n T \right)^2 - 3\left( \sin(\pi n T) - \pi n T \cos(\pi n T) \over \pi n T \right)^2,$$
where $T$ is the linear detrending time.

3.4.2 Experimental transfer functions

A series of schematics with 9 parts is presented in (figure 3.2) which summarizes the explanation of the experimental determination of the high frequency transfer function correction. The part in concern will be denoted as number in this section.

Before starting the spectral correction procedure, the model co-spectra (2.46 and 2.47) has to be solved by determining a stability dependency for the frequency with the maximum contribution to the flux. Horst (1997) shows that the correction
Table 3.4: Variables needed for the theoretical definition of the high frequency transfer function. $U$ – mean wind speed, $s_{irga}$ – sensor separation between IRGA and sonic anemometer, $n$ – frequency.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Explanation</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_{CP}$</td>
<td>Time constant, closed-path</td>
<td>0.109 s</td>
</tr>
<tr>
<td>$\tau_{OP}$</td>
<td>Time constant, open-path</td>
<td>0.032 s</td>
</tr>
<tr>
<td>$\tau_{sonic}$</td>
<td>Time constant, sonic</td>
<td>0.016 s</td>
</tr>
<tr>
<td>$s_{CP}$</td>
<td>Sensor separation, closed-path</td>
<td>0.12 m</td>
</tr>
<tr>
<td>$s_{OP}$</td>
<td>Sensor separation, open-path</td>
<td>0.21 m</td>
</tr>
<tr>
<td>$p$</td>
<td>Sonic path length</td>
<td>0.175 m</td>
</tr>
<tr>
<td>$\Lambda$</td>
<td>Tube length (closed-path)</td>
<td>40 m</td>
</tr>
<tr>
<td>$d$</td>
<td>Inner diameter of the tube</td>
<td>8 mm</td>
</tr>
<tr>
<td>$v$</td>
<td>Flow rate in the tube</td>
<td>171/min = 5.64 ms$^{-1}$</td>
</tr>
<tr>
<td>$\nu_{air}$</td>
<td>Kinematic viscosity of air</td>
<td>$1.5 \cdot 10^{-5}$ m$^2$s$^{-1}$</td>
</tr>
<tr>
<td>$Re = vd/2\nu_{air}$</td>
<td>Reynolds’ number</td>
<td>$\approx 3750$</td>
</tr>
<tr>
<td>$f_s = ns_{irga}/U$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$f_p = np/U$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The factor in equation 3.52 converts to

$$CF = \frac{w's'_{meas}}{w's'} = \frac{1}{1 + 2\pi n_m \tau_s}$$ (3.6)

assuming a co-spectral form

$$\frac{nC(n)}{w's'} = \frac{2}{\pi} \frac{n/n_m}{1 + (n/n_m)^2}.$$ (3.7)

Now a value $\tau_s = 0.6$ (and the corresponding $TF_{HF}$) has to be assumed for solving the frequencies $f_m$ from

$$f_m = \frac{1 - CF}{CF2\pi \frac{U}{z-z_0} \tau_s}.$$ (3.8)

Here the correction factor ($CF$) is calculated by integrating the measured normalized temperature co-spectra multiplied by the transfer function. Next, the coefficients ($a_1, a_2, a_3$) in equation 2.48 can be determined by fitting a curve as a function of stability and the model co-spectra can be used. [2]

The procedure of determining a function for $\tau_s(RH)$ starts with a careful data selection, [3]. The data for the calculations have to meet the following criteria:
April-September, daytime, $H > 50 \text{ Wm}^{-2}$, $-0.01 < \zeta < -0.00001$, wind from either the vegetation or the road sector. In addition, a visual inspection is done for the temperature and scalar co-spectra. A typical error causing the omittance of data based on visual inspection is white noise, which occurs at high frequencies. It is caused either by sensor noise that rises above the signal level or by sporadic spikes in the signal caused by measurement malfunctions. White noise is seen as a rise in the co-spectra in the high frequency end. The slope approaches typically unity in a $nC(n)$ plot (Kaimal and Finnigan, 1994). Half an hour data are used and the selected data records are divided into seven relative humidity classes with an equal amount of data records. The following calculations are applied to each RH class separately.

The high frequency transfer function used for the closed-path IRGA can be empirically defined as (Aubinet et al., 2000),

$$TF_{HF}(n) = \frac{N_T C_{ws}(n)}{N_s C_{wT}(n)},$$

(3.9)

where the normalization factors are unattenuated covariances

$$\frac{N_T}{N_s} = \frac{\int_{n_1}^{n_2} C_{wT}(n)dn}{\int_{n_1}^{n_2} C_{ws}(n)dn}.$$  

(3.10)

Here $n_1$ and $n_2$ are limit frequencies (Hz). $n_1$ has to be low enough for the integral to take into account a sufficient amount of contribution from the low frequency and high enough that the scatter at the low frequency end does not disturb the calculations. $n_2$, on the other hand, has to be high enough for the integral to be sufficient and low enough for the attenuation to be negligible. Here the limits were chosen to be 0.004 Hz and 0.8 Hz. Equation 3.9 now corresponds to the whole set of terms in equation 3.5. First, the RH class averages of $C_{ws}(n)/N_s$ and $C_{wT}(n)/N_T$ are calculated. Second, the values are binned into 76 logarithmically evenly spaced points. Third, the high frequency transfer function is determined from 3.9. Finally, equation 2.53 is fitted to $TF_{HF}(n)$ to get the response time of the measurement system $\tau_s$. The procedure is repeated for all seven relative humidity classes and a value for $\tau_s$ is gained for each class. A curve of the form

$$\tau_s = a_1 + a_2 \cdot RH^{a_3}$$

(3.11)

is fitted to the response time data.
The experimental definition does not contain the effect of dynamic response of the sonic, the scalar path averaging and the sensor response mismatch. In other words, the temperature measurements are assumed to be measured accurately and the co-spectra of temperature is used as the real model spectra. Thus, local spectral similarity between water vapor and temperature is assumed.

Now that $\tau_s(RH)$ has been successfully determined, one can move to making the actual corrections to the measured covariances. First $TF_{HF}$ has to be calculate for each time point using the RH data. In addition, $TF_{LF}$ according to equation 3.6 is calculated, but it is constant for all time points. The model co-spectra with the stability dependent $f_m$ is also needed for substituting $C_{ws}(n)$ in equation 2.52. Here global spectral similarity is assumed meaning that the normalized co-spectrum of potential temperature and scalars have the same shape independent of the measurement location. The correction factor can now be defined as,

$$CF = \frac{\int_0^\infty TF(n) \cdot C_{w\theta, model}(n) dn}{\int_0^\infty C_{w\theta, model}(n) dn}.$$  

The denominator is unity according to its definition. Finally, the integrations in equation 3.12 are done using the Composite Simpson’s method, 8. In the method the data is split up to $n$ sub intervals, where $n$ must be an even number.

$$\int_a^b f(x) dx \approx \frac{h}{3} \left[ f(x_0) + 2 \sum_{j=1}^{n/2-1} f(x_{2j}) + 4 \sum_{j=1}^{n/2} f(x_{2j-1}) + f(x_n) \right],$$  

where $h$ is the step length, given by $h = (b-a)/n$. After defining $CF$, the calculated covariances can be corrected ($w's_{\text{cor}} = \frac{w's_{\text{meas}}}{CF}$), and the other corrections can be applied and the flux can be calculated, 9.
A model co-spectra is used in
\[ C_{\text{model}} = \frac{nC_{ws}}{w'S'} = \frac{a_1 f / f_w}{(1 + a_2 f / f_w)^2} \]

1. **INITIATION**
   - Determine \( f_m(t) \) assuming \( \tau_s = 0.6 \).
   - \( f_m = \frac{1-CF}{CF2n(z-d)/U + \tau_s} \)

2. **CALCULATING \( f_m \)**
   - \( f_m \) is a constant in the unstable range.
   - \( f_m = 1 - \frac{TF_{low}}{CF} \frac{p_U}{U(z-d)} \)

3. **(RH)**
   - Find half an hour data records that fulfill the criteria:
     - April-September, day-time
     - Unstable
     - Vegetation or road sector
     - Visual inspection of the co-spectra
   - Divide the half an hour data records into 7 classes with an equal amount of time points.

4. **MAKE CALCULATIONS TO EACH RH CLASS**
   - Calculate average
   - Bin into logarithmically evenly spaced 76 points
   - Calculate \( CF = \frac{\int TF_{low} C_{\text{model}} dn}{\int C_{\text{model}} dn} \)

5. **FIT CURVE TO SOLVE \( \tau \)**
   - \( TF_{\text{high}} = \frac{1}{1 + (2\pi \tau_s)^2} \)
   - Repeat for all RH classes

6. **NOW THERE IS ONE VALUE OF \( \tau \) PER EACH RH CLASS. FIT CURVE TO DATA AND GET \( n(RH) \).**

7. **CORRECTION**
   - Go through each measurement time and calculate:
     - \( \tau_s \) and \( TF_{\text{low}} \) from RH data
     - \( TF_{\text{low}} \) from Rannik and Vesala 1999
     - model co-spectra

8. **MULTIPLY THE MODEL CO-SPECTRA BY THE TRANSFER FUNCTIONS. INTEGRATE NUMERICALLY TO GET THE CORRECTION FACTOR.**
   - \( CF = \frac{\int TF_{\text{low}} C_{\text{model}} dn}{\int CF_{\text{model}} dn} \)
   - \( \frac{w'S'}{w'S_{\text{corr}}} = \frac{w'S_{\text{model}}}{CF} \)

Figure 3.2: A series of figures explaining the experimental transfer function correction. Local spectral similarity is assumed in [4] and global spectral similarity in [8]. In addition, a value for \( \tau_s \) is assumed in [1].
3.5 A wavelet-based spectral correction

In the following a new type of a method to correct the loss of high frequency variation is represented. It is based on the manipulation of raw data and thus does not use the concepts of a transfer function or response time. Furthermore, the mixed velocity scalar structure function for assessing the performance of the wavelet-based correction is represented. A case study for water vapor will be used to clarify the procedure. The data is from ideal conditions in 5th of June 2008, 11:00-11:30. At this time, the wind was blowing from the vegetation sector and the atmosphere was unstably stratified, $\zeta = -0.26$. The stationarity criteria of the fluxes are also fulfilled. A data record of length $2^{14} = 16384$ is used.

3.5.1 Correction procedure

The correction starts with calculating the power spectrum of the scalar of interest ($\text{H}_2\text{O}$) using wavelet analysis. The power spectrum with its 14 points is shown in figure 3.3 a) with the model power spectrum calculated for SMEAR III (equation 4.1). The peak frequency $f_m$ is determined by fitting a curve to each half an hour data record and finding the maximum. Amplitude attenuation at high frequencies is apparent. Now the goal is to adjust the amplitudes at the high frequency end using the undamped amplitudes adjacent to them (marked with a dashed ellipse in figure 3.3 a)). Before doing so, the existence of white noise shown in figure 4.2 has to be accounted for.

White noise alters the phase of variation of data and tube attenuation dampens the amplitudes. Consequently, if the dampened amplitudes are wished to be restored to their real magnitude using the information on the phase, the data should not be contaminated by white noise. A method to check if there is too much white noise was introduced by Donoho and Johnstone (1994). If it is assumed that all wavelet coefficients at the finest scale ($Wf^{(j=1)[p]}$) are contaminated by white noise, the maximum amplitude of white noise is

$$WN = \sqrt{2 \log(N/2)} \cdot \sigma_{s=1}.$$  \hspace{1cm} (3.14)

Here $\sigma_{s=1}$ is the standard deviation of the wavelet coefficients at the finest scale. Now all absolute amplitudes of wavelet coefficients from other scales can be compared.
with this value. If the absolute amplitude at scale index \( j \) and position index \( p \) fulfils
\[
\frac{|Wf^{(j)}[p]|}{\log(2)} > WN,
\]  
(3.15)
the wavelet coefficient is considered uncontaminated. If 80% of coefficients at a certain scale are uncontaminated, the scale can be used in the spectral correction. This procedure gives a vector of the length \( J \) with percent values (figure 3.4). A single value assessing the effect of white noise of a data record can also be derived. The signal-to-noise ratio ([Donoho and Johnstone] 1994)
\[
SN = \frac{\sigma_s}{WN},
\]  
(3.16)
where \( \sigma_s \) is the standard deviation of the time series of the scalar. If \( SN > 3 \), the data is acceptable. Now there are two measures of white noise: one for each scale separately and another representing the whole data record. The correction is not done if \( SN < 3 \). For the case study data, \( SN = 10.03 \).

Now that it has been ensured that the signal-to-noise ratio is high enough, the points used for interpolation can be chosen. First, the model power spectrum, which depends on \( U \), \( \zeta \) and \( z \), is used to find the scales in the inertial subrange. A slope of \( nS(f) \) in a logarithmic scale is calculated using different amount of points from the small scale end. The slope is calculated for points \( m < j < J \) where \( m[2, J-4] \). The largest scale that still gives \( -1.67 < d \log(nS)/d \log(f) < -1.65 \) is taken as the first point in the inertial subrange. This data point together with other points for
This page contains a description of a method for calculating the slope of a power spectrum and interpolating wavelet coefficients. The method involves the following steps:

1. Calculating the slope of the power spectrum using wavelet coefficients. The slope is calculated as \( d \log(nS)/d \log(f) \) for the measured power spectrum, where \( n \) is the scale index.
2. Evaluating the fraction of uncontaminated wavelet coefficients as a function of scale index (equation 3.15). This is done by selecting at least 3 points that fulfill the limit of 80% of uncontaminated wavelet coefficients. If the slope calculated from these points is within the limits [-1.8, -1.5], the linear fit will be used for interpolation. Otherwise, a -5/3 slope according to Kolmogorov will be used. Scales with indices 4 to 7 are used in the case study for linear fitting (figure 3.3).

When the points used for interpolation are chosen, the interpolation and amplitude adjustment can be conducted. First, unattenuated amplitudes for the 4 smallest scales \((j = 1, 2, 3, 4)\) are made using interpolation. Figure 3.3 a) shows the squares of the adjusted amplitudes with red circles. The wavelet coefficients of the four last scales are multiplied by the ratio between the unattenuated and attenuated amplitudes

\[
Wf^{(j)}[p] = Wf^{(j)}[p] \cdot \frac{A}{A_{\text{atten.}}}, \quad j = 1, 2, 3, 4.
\]

Note that the wavelet coefficients include the variation in time and thus the time-localization feature of the wavelet transfer is essential here. Now the time series can be reconstructed using the adjusted wavelet coefficients and making an inverse wavelet transform. A small window of the original and adjusted time series of \( q' \) is shown in figure 3.5. The adjusted time series is well constructed and has clearly an addition of small scale variation. The co-spectra calculated from the original time series and the manipulated time series are shown in figure 3.3 b). Also, the model spectra by Kaimal et al. (1972) is shown. The attenuation of small scale variations is accounted for in the adjustment process.
variation of the scalar has been successfully corrected and the co-variance at high frequencies is lifted up to obey the \(-4/3\) dependency. The co-variance grew in this case 1.5%. Note that the correction factor for the transfer function correction with a RH dependency was in this case 0.91. Note that no spectral similarities were used here and the vertical wind speed data was not altered at all. The model power spectrum was only used for defining the scales in the inertial subrange.

The low frequency correction (equation 3.6) can be applied to the co-spectra calculated from the adjusted time series. After that, the correction factor can be calculated by numerically integrating the co-spectra and finally the flux value from the original time series can be corrected. The numerical integration of only \(J = 14\) points is prone to errors, but the integration is needed if the low frequency spectral correction is also desired. The correction factor can also be directly calculated from the covariances of the original and amplitude adjusted time series, but in such a case the low frequency correction is not taken into account.

### 3.5.2 Mixed velocity scalar structure function

An independent measure called the mixed velocity scalar structure function \((D_{ssu})\) is used to assess the added small scale variation in the adjusted time series. Structure parameters are widely used measures of the structure of small-scale turbulent variation (e.g. [Kaimal and Finnigan](1994), p. 51). The mixed velocity scalar
structure function is defined as (e.g. Katul et al. (1997))

\[ D_{ssu}(r) = (s(x + r) - s(x))^2 \cdot |u(x + r) - u(x)|, \]  

(3.18)

where \( s \) is the scalar, \( u \) the horizontal wind speed, \( x \) denotes a data point and \( r \) is the separation distance between two data points. \( D_{ssu} \) can also be derived from the Navier-Stokes equation assuming locally isotropic turbulence, \( r >> \eta \) and that \( \nu \) is small. The assumptions are justifiable for air and the result is

\[ D_{ssu}(r) \simeq -\frac{4}{5} N_s r, \]  

(3.19)

where \( N_s \) is one half of the scalar dissipation rate analogous to the energy dissipation rate \( \epsilon \) mentioned earlier. Now it can be seen that if a turbulence data record follows Kolmogorov similarity \( D_{ssu} \) is linearly dependent on \( r \). In other words, the slope \( S_{D_{ssu}} = d \log(D_{ssu}(r))/d \log(r) \) should equal one. If the slope is below one, the power spectrum of the scalar falls too fast in the high frequency range, that is, it is attenuated. If the slope exceeds unity the the power spectrum falls too slow. The slope is calculated for an arbitrary range of \( r [1,10] \). Changing the range to \( [1,20] \) does not have an effect. \( S_D \) is calculated for the adjusted time series and the original time series to resolve the effect the correction had on the small scale variation. The test is independent of any corrections and uses the horizontal wind speed instead of the vertical wind speed. The test will be used to assess the performance of the wavelet-based spectral correction.

\footnote{The Kolmogorov microscale: \( \eta = (\nu^3/\epsilon)^{1/4} \)}
Results and discussion

In this section, complementary meteorological data will be first briefly introduced. Next, the spectral correction results will be presented. First, the average spectra are shown and local spectral similarity is assessed. Second, the results of the transfer function and wavelet-based spectral corrections. Finally, the extensive time series of the components in the energy balance are discussed. The focus will be on the residual term and a literature review concerning it will be shown.

4.1 Meteorological background data

The meteorological data for temperature, relative humidity, pressure, wind speed and rain are in figure 4.1. The gray dots are half an hour averages and the black lines the 5-day running means, and data from the whole measurement period (Dec 2005-Jun 2009) are displayed. The running mean of temperature ranges from -12 to +21 °C and winter 2007 was the coldest whereas winter 2008 was the warmest. The annual averages for 2006-2008 are 6.7, 7.0 and 7.4 °C which all exceed the climatological average of 5.6 °C. The running mean of relative humidity varies between 40 and 92%. The maximum is observed in winter whereas the driest times are in early summer. The annual averages are all close to 74.5%. The synoptic scale variation in pressure is clear and the years do not differ much from one another. Wind speed is highest during winter and it may even exceed 10 m/s. The annual averages are 3.4, 3.5 and 3.7 m/s. The annual precipitation varies more from year to year as the yearly sums are 533, 693 and 812 mm. The climatological mean is 642 mm, as mentioned earlier. The annual means of 2009 are not reported since a full data record is not...
yet available.

Figure 4.1: Basic weather variables in 2006-2009. Gray dots are half an hour values and black lines 5-day running means.
4.2 Spectral corrections

4.2.1 Average power spectra and co-spectra of temperature and water vapour

Figure 4.2 displays the average power spectrum in unstable situations at SMEAR III. Variance of water vapor is clearly lost at the high frequency end. The $-2/3$ slope in the inertial subrange according to Kolmogorov is changed into $-8/3$ due to the additional dependence on $f^{-2}$ (eq. 2.53). Also white noise is present for both water vapor and temperature data. The data is severely contaminated at the last high frequencies which appears as a 1:1 line. In addition, scatter and elevated variance values at low frequencies can be seen.

The high frequency attenuation of water vapor is naturally also seen in the co-spectrum (figure 4.3). Here the $-4/3$ slope is converted to $-10/3$ due to the additional dependency on $f^{-2}$. Scatter is present in the low frequency end, but the values are not elevated as in the power spectrum. Actually the standard high frequency correction will act to increase the values. In addition, remainders of white noise can also be seen at the highest frequencies.

Figures 4.2 and 4.3 display also the fits for temperature and H$_2$O power and co-spectra in unstable stratification. Only points 7-50 out of 75 were used in the fit to avoid the effect of contamination in both frequency ends. The fit at $f \geq 1$ for $wT$ co-spectra is not shown due to the attenuation. Power spectral peaks for $T$ and H$_2$O are 0.683 and 0.320 whereas the co-spectral peak of $T$, $f_m=0.11$, is used for both scalars. The integrals of the fits equal one, but the transition point at either $f = 0.15$ or $f = 1$ is not continuous. The fits for the two scalars give about the same coefficients for the first frequency domain whereas the second fit domain is badly distorted for the water vapor power spectrum. In addition, the widely accepted co-spectral model is shown (equation 2.46). The roll-off for the model co-spectra is a bit faster at high frequencies but the model is anyhow used in the transfer function correction. The corresponding power spectral model (equation 2.50) did not fit to the data at all and is thus not shown. A continuous power spectral model for SMEAR III

\[
\frac{fS_q(n)}{\sigma_q^2} = \frac{0.974 f/f_m}{(1 + 1.767 f/f_m)^{5/3}} \tag{4.1}
\]
is used in the wavelet-based spectral correction instead of the piecewise determined function.

The spectra for stable situations depend on stability, as mentioned before, and average spectra are not shown due to the limited amount of stable situations. The frequency with the maximum contribution to the co-spectra of temperature can be defined by an equation of the form \( f_m = a_1 (1 + a_2 \zeta a_3) \), when \( L > 0 \). The coefficients for SMEAR III are in table 4.2.1 with the corresponding values for SMEAR II (Rannik et al., 2004). The results are quite similar between the two sites. A similar fit for \( f_m \) for power spectra in stable stratification was not done. The reason is the

\[
 f_m = n(z - z_d)/U
\]

is the normalized frequency and it is further normalized by its value at the spectral peak, \( f_m \) (0.068 for temperature and 0.032 for water vapor). The slopes \(-2/3\), \(-8/3\) and \(1:1\) are for Kolmogorov theory, attenuated spectrum and white noise, respectively. Data is for unstable situations and times when the wind is blowing from the vegetation sector.

Figure 4.2: Average normalized power spectra of temperature and water vapor. The normalization is done with the standard deviation (\( \sigma \)) of the variable. \( f = n(z - z_d)/U \) is the normalized frequency and it is further normalized by its value at the spectral peak, \( f_m \) (0.068 for temperature and 0.032 for water vapor). The slopes \(-2/3\), \(-8/3\) and \(1:1\) are for Kolmogorov theory, attenuated spectrum and white noise, respectively. Data is for unstable situations and times when the wind is blowing from the vegetation sector.
Figure 4.3: Normalized co-spectra of temperature and water vapor. The normalization of the heat and water vapor fluxes is done with the covariances $w^T$ and $w^q$, respectively. $f \equiv n(z - z_d)/U$ is the normalized frequency and it is further normalized by its value at the spectral peak, $f_m = 0.11$ for both co-spectra. The model co-spectra is from equation 2.46. The slopes $-2/3$, $-8/3$ and $1:1$ are for Kolmogorov theory, attenuated spectrum and white noise, respectively. Data is for unstable situations and times when the wind is blowing from the vegetation sector.

Vesala et al. (2008) analyzed power and co-spectra from two time periods from SMEAR III. The data represented the urban and the vegetation sector and $u$, $T$ and CO$_2$ were studied. The power and co-spectra are generally in agreement with the results here, but naturally the averages represented here are smoother. They also found the effect of a spectral shortcut as a step in the power spectra of $T$ and CO$_2$ and in the co-spectra for all variables for the urban sector. The shortcut was at frequencies corresponding to eddy sizes similar to building dimensions. This phenomenon is not studied in this work. Vesala et al. (2008) did not detect the peak lack of power spectra that fulfill the criteria of stratification and pass the overall visual inspection.
Table 4.1: Coefficients for equation 2.48 determining the co-spectral peak for temperature.

<table>
<thead>
<tr>
<th>$a_1$</th>
<th>$a_2$</th>
<th>$a_3$</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.11</td>
<td>2.35</td>
<td>0.33</td>
<td>Urban</td>
</tr>
<tr>
<td>0.11</td>
<td>0.95</td>
<td>0.00</td>
<td>Road</td>
</tr>
<tr>
<td>0.11</td>
<td>1.99</td>
<td>0.38</td>
<td>Vegetation</td>
</tr>
<tr>
<td>0.9</td>
<td>4.5</td>
<td>0.78</td>
<td>Rannik et al. (2004)</td>
</tr>
</tbody>
</table>

...shifting of spectra due to the wake effect that was introduced earlier (section 2.5.5).

4.2.2 Local spectral similarity between scalars

A case study of local spectral similarity will be presented in the following. The same data record as in the case study used for clarifying the wavelet correction procedure in section 3.5 will be used. Local spectral similarity is assumed in the transfer function spectral correction and is thus assessed here. Global spectral similarity will not be tested for SMEAR III.

Figure 4.4 shows a wavelet half-plane of the relative transport efficiency of turbulence for heat and water vapor ($w' s'/\sigma_w \sigma_s$). The representation takes full benefit of the time-localization feature of wavelet analysis: time is on the x-axis and scale on the y-axis. The higher resolution at small scales mentioned earlier is now evident. Note also that the number of scales displayed on the y-axis is $J = 14$.

Several features can be pointed out from the half-plane. First of all, the existence of coherent structures is evident: the turbulent transfer of both scalars consists of episodic eddy groups where small swirls are superimposed on the larger eddies. Small scale turbulence do not have organized structures but as the scale increases the turbulence tends to organize into discrete structures. Second, the high frequency attenuation of water vapor can be seen as lighter shades at the small scales. The four smallest scales that are amplitude adjusted in the wavelet correction are marked with a dashed rectangle. Furthermore, weaker H$_2$O transport relative to heat transport is observed throughout the time period and one episode is denoted by a triangle. When the squares of the wavelet coefficients of $w'T' / \sigma_w \sigma_T$ and $w'q' / \sigma_w \sigma_q$ are summed over time per each scale, it turns out that the transport of temperature is much more efficient for large eddies (not shown). Also the transport efficiency for small scales...
was larger for heat than H$_2$O, but it might be due to the attenuation of small perturbations. To the contrary, an episode with stronger H$_2$O transport can also be found (rectangle). Even a case with opposite fluxes can be found (ellipse).

Figure 4.4: A wavelet half-plane for heat and water vapor flux on 5th June 2008, 11:00-11:30. The colors indicate the relative transport efficiency of turbulence for temperature and water vapor fluctuations. The negative values are for cases with negative covariance, that is, the flux is directed downwards. The time is on the x-axis and the scale of the turbulence on the y-axis. Dashed triangle – smaller transport efficiency of water vapor, ellipse – fluxes to opposite directions, rectangle – transfer efficiency of water vapor stronger, triangle – transfer efficiency of heat stronger. See text for further explanations of the shapes.

Figure 4.5 depicts the co-spectra of water vapor and heat in addition to the sum of the co-spectra and its standard deviation over all position indices (equation 2.29). The co-spectra seem to coincide quite well for all but the four smallest scales, which were corrected in section 3.5. More discrepancy is observed in the co-spectra.
added by its standard deviation. The values for heat are always higher than those for water vapor. The difference is an indication of larger variability in the heat flux which was also observed as a higher transport efficiency in the half-plane figure.

Moriwaki and Kanda (2006) made a similar half-plane analysis of the dissimilarity of scalars over a suburban area. They also pointed out a stronger transfer efficiency of heat relative to H$_2$O and they argued the reason to be heterogeneity in sources and the different physical properties of scalars. Heat is an active scalar that produces the thermal turbulence itself whereas water vapor is passive in that sense. (Katul and Parlange, 1994) also used Wavelet analysis and showed a dissimilarity between water vapor and heat at a bare soil field for large wave numbers. They showed that variability of energy in space compared to the mean was very large for heat and that heat behaves in an anisotropic way again emphasizing the active role of the scalar. They suspect the dissimilarity to stem from non-uniform heating and cooling at the surface which influences the active role of temperature in the dynamics of small-scale turbulence.

Figure 4.5: The co-spectra of heat and water vapor in addition to the co-spectra plus its standard deviation (equation 2.29) on 5th of June 2008, 11:00-11:30. Note that $\sigma_{nC/cov_{ws}}$ is zero for the largest scale because there is only one coefficient available for the largest scale.

4.2.3 Transfer function correction and effect on LE

The experimentally defined transfer function (eq. 2.53) for water vapor varies from year to year. Transfer functions per year without relative humidity separation of data are shown in figure 4.6. The response time is 0.39 s in 2007 and grows to
0.49 s in 2008. The increase is due to tube ageing and a reduction in flow rate. In summer 2007 the flow rate in the inlet was lower than desired. Also, in fall 2008, the flow rate in the sampling tube gradually decreased and the problem was fixed in winter 2008. In 2009 the response time has dropped to 0.29 s which can be partially explained by the increase of flow rate compared to 2007 and 2008. The theoretical transfer function for closed-path $\overline{w'q'}$ is also displayed in the same figure. Clearly, the theory gives too high values for the transfer function meaning an underestimated response time. Aubinet et al. (2000) also came to the conclusion that the theoretically determined response times are inadequate for high frequency spectral corrections. Furthermore, the theoretical transfer functions for $\overline{w'u'}$, $\overline{w'T'}$ and $\overline{w'q'}$ open-path are shown (equation 3.5). The transfer function for covariances based on sonic data only are naturally much higher because the response time is low since air samples are not transported anywhere.

Figure 4.6: Transfer functions and corresponding response times for water vapor for 2007-2009. Note that only data between frequencies 0.004 and 0.8 Hz were used and the transfer function had to be [0, 1.3]. No relative humidity classes were used here. Also the theoretical $TF$ for $\overline{w'u'}$, $\overline{w'T'}$, $\overline{w'q'}$ open-path and $\overline{w'q'}$ closed-path are shown (equation 3.5).
The response time for H$_2$O clearly depends on relative humidity. Figure 4.7 shows the response time as a function of relative humidity for years 2007-2009. The data were binned to seven relative humidity classes each consisting of an equal amount of data points, and the response time for each class was calculated. A non-linear fit was done to the response time data and is also shown in figure 4.7. The general level of the response time is consistent with the average values obtained without relative humidity binning. Year 2007 has the strongest relative humidity dependency with values $\tau(30\%) = 0.35$ s and $\tau(100\%) = 5.0$ s. For 2008 the corresponding values are 0.34 s and 1.76 s whereas for 2009 they are 0.15 s and 1.4 s. Similar relative humidity dependency was neither observed for carbon dioxide nor particles.

Similar relative humidity dependency of the response time has also been detected in recent studies. Ibrom et al. (2007) found that the cut-off frequency of water vapor decreased exponentially with increasing relative humidity in a Beech forest in Denmark. Mammarella et al. (2009) made a similar analysis as here for 6 years of data from the SMEAR II station in a Scots pine forest in Finland. Their values for the non-linear fit shown in figure 4.7 were of the same order of magnitude. They also assessed the effect of tube age by changing the sampling tube to a new one. With a new clean tube, the relative humidity dependency of $\tau$ was very small and heating of the tube was able to minimize the water vapour absorption on tube walls. To the contrary, the relative humidity dependency increased with tube age or tube dirtiness. Leuning and Judd (1996) and Su et al. (2004) also found that tube age...
increases the response time for water vapor, but neither one assessed the effect of relative humidity. The latter study was made for data from a mixed hardwood forest in mid-western USA. Massman and Ibrom (2008) made a new type of a model describing tube attenuation and treating water vapor as a non-passive tracer. They also took into account relative humidity variation, but the model results were not successful: the attenuation seemed to increase as a function of the Reynolds’ number. Clearly, more theoretical work is needed to resolve the reasons for water vapor fluctuation attenuation.

Latent heat fluxes without any spectral correction give generally smaller values than spectral corrected fluxes. The fluxes grow on average 18, 15 and 16% for years 2007-2009 when a constant response time is used in corrections. Moreover, when $\tau$ depends on $RH$, the corresponding values are 21, 17 and 14%, respectively. Apparently, the spectral correction with a relative humidity dependency has more impact on the flux in 2007 and 2008 compared to a correction without a relative humidity dependency. To the contrary, the opposite is observed for 2009. The correction is larger the larger the flux itself (not shown) and more significant corrections are always observed during night-time and in winter.

The relative corrections observed here are congruent with other similar studies. Su et al. (2004) calculated $LE$ corrections of 10% for newer tubes and 20% for older tubes. Ibrom et al. (2007) got as high an increase as 42% for monthly latent heat fluxes. Moreover, Mammarella et al. (2009) calculated a 7% correction for new tubes and a 10-40% correction for an old tube. Note that all of these studied were conducted on the basis of data from a forest site.

The data from SMEAR III also allow the comparison between latent heat fluxes measured with an open- and a closed-path analyzers. $LE$ measured with an open-path IRGA is larger on average than the un-corrected $LE$ measured with the closed-path IRGA ($LE_{OP}$ and $LE_{CP}$ in figure 4.2.3). The annual average differences are 1.0, 3.8 and 5.8 Wm$^{-2}$, for 2007-2009. When $LE_{CP}$ is corrected using a constant response time, the differences are -4.0, -3.1 and 3.6 Wm$^{-2}$ meaning that $LE_{CP} > LE_{OP}$ for 2007 and 2008. When the relative humidity dependency of $\tau$ is taken into account, the values are -4.1, -1.7 and 5.5 Wm$^{-2}$. In other words, the average difference decreased from 3.5 to -0.1 Wm$^{-2}$ when comparing uncorrected fluxes with spectral corrected fluxes using $\tau(RH)$. The open-path data has not been corrected according to Burba et al. (2008). This additional correction would elevate the high $LE$ values.
of the open-path IRGA and thus lead to better correspondence between the open- and closed-path analyzer. It must be also noted that open-path data is poor in high relative humidity conditions and thus has been neglected. This of course affects the current comparison. Also it must be kept in mind that the data coverage is one year only in 2008 and a half a year in 2007 and 2009.

4.2.4 Initial results of a wavelet-based spectral correction

The wavelet-based spectral correction was applied for now only to data from times when the wind was blowing from the vegetation sector and the atmosphere was unstably stratified. The reason for the selection is the laborious determination of the model power spectra. The dependency \( f_m(\zeta) \) for stable conditions could not be determined for SMEAR III because of the scarce occurrence of stable situations. In future, the analysis may be extended to other time points, too.

Before going into the correction factor results, the raw data manipulation method will be assessed. The slope of the logarithm of the mixed velocity scalar structure function, \( S_{D_{ssu}} \) is unity if the Kolmogorov similarity applies (section 3.5). Figure 4.9 displays \( S_{D_{ssu}} \) as a function of relative humidity for original and manipulated data. A significant non-linear dependency is observed for the original data, but the dependency is clearly weakened after the manipulation. Especially the number of values \( S_{D_{ssu}} < 0 \) (indicating attenuation) has decreased substantially as a result of the high frequency amplitude adjustment. This is precisely what was intended.

The correction factor of the wavelet-based correction (\( CF_W \)) is much larger than that of the transfer function correction (\( CF_{TF} \)). This indicates that the wavelet-based method does not increase the fluxes as much as the traditional transfer function method. Figure 4.10 shows the normalized occurrence of different CF values for the transfer function correction without (\( CF_{TFc} \)) and with (\( CF_{TFc(RH)} \)) a relative humidity dependency in addition to \( CF_W \). The Wavelet correction factor is here calculated directly from the time series as a covariance and thus does not include the low frequency correction. The two transfer function based corrections give relatively close values though the correction factor with the RH dependency is a bit smaller (larger fluxes), as mentioned earlier. \( CF_{TFc} \) and \( CF_{TFc(RH)} \) are both always below unity according to their definition. \( CF_W \) has much higher values with a mean of 0.994 and a minimum of 0.870 and a maximum of 1.084. The larger than
Figure 4.8: A comparison of latent heat fluxes. $LE_{OP}$ – latent heat flux measured with the open-path IRGA, $LE_{CP}$ – latent heat flux measured with the closed-path IRGA, $LE_m$ – $LE_{CP}$ without spectral corrections, $\tau_{H_2O}=\text{const.}$ – response time does not depend on relative humidity, $\tau_{H_2O}(\text{RH})$ – response time depends on relative humidity according to the equations in figure 4.7. Gray lines are 1:1 and red lines the linear fits.
Figure 4.9: Slope of log($s'/s''$) versus log($r$) as a function of relative humidity. Left panel shows original data and right panel the amplitude adjusted data. If the value is below unity, the decay is too fast at high frequencies. The opposite holds for values above unity. RH in the fit is [0,1].

Unity results are possible if the high frequency decay has been slower than the -5/3 slope predicted by the Kolmogorov theory. For $CF_{TF}$ with a RH dependency, the mean, minimum and maximum are 0.856, 0.520 and 0.971. The discrepancy implies a larger correction to the fluxes by the transfer function method compared with the wavelet-based method. The reason may be that the wavelet-based correction only corrects the attenuation in the inertial subrange whereas the transfer function correction acts on a larger frequency range.

The physics behind the correction factor from the wavelet-based correction can be assessed using basic correlation analysis. There is a weak linear dependency between $S_{D_{ss}}$ and $CF_W$ ($R = -0.37$) implying that the covariance is underestimated when the high frequency end of a scalar is attenuated. This is precisely the case. A surprisingly weak positive correlation between $CF_W$ and $CF_{TF,\tau(RH)}$ is also observed ($R = 0.26$). In addition, wind speed is the only environmental variable that shows some systematic behaviour when set against $CF_W$. A negative linear dependency ($R = -0.30$) was observed. Other variables like stability and relative humidity did not seem to exhibit a dependency of any form.
Figure 4.10: Normalized distribution of the correction factor for three different correction methods: 
\( \tau = \text{const.} \) – transfer function method without a relative humidity dependency, 
\( \tau(RH) \) – transfer function method with a relative humidity dependency, Wavelet – wavelet-based spectral correction. Data is shown only for time points for which all correction factors are available.

4.3 Urban energy fluxes

4.3.1 Radiation

The components of net radiation (equation 2.55) are measured at one fixed point and thus the upward fluxes represent the characteristics (albedo and temperature) of the specific circular area under the sensor (section 3.2). As stated before, the area at SMEAR III consists of grass, shrubs, rock and trees. Consequently, the interesting phenomena related to a reduced sky factor, among others, are not detectable at SMEAR III (section 2.5.1). What could be seen is an increase of LW \( \downarrow \) owing to the heat island effect or an increase in SW \( \downarrow \) owing to a larger number of scattering particles in the air. The detection of these two phenomena is not restricted by the current location of the sensor but by the lack of comparative measurements from a rural site near Helsinki. Thus, the comparison will not be done.

The almost four year time series of all the components contributing to net radiation are shown in figure 4.11. The behavior of the downward directed short-wave radiation essentially reflects the existence of cloud cover and the declination angle of the sun. Its 5-day running mean shows a peak during summer (380 Wm\(^{-2}\), a
minimum in winter \((30 \text{ Wm}^{-2})\) and a synoptic scale time variation of a couple of days. The hour-to-hour data show maximums as high as \(830 \text{ Wm}^{-2}\) during summer days and zero values during night. The monthly average diurnal variation reveals the cloud cover effect as the declination angle is essentially the same. In July 2006, a month with fair weather, \(SW \downarrow\) is \(670 \text{ Wm}^{-2}\) at noon whereas the corresponding value for July 2007, a rainy month, is only \(490 \text{ Wm}^{-2}\) (not shown). The upward directed short-wave radiation is just the albedo times \(SW \downarrow\) (equation \(2.55\)). Thus, the data reveals the times of altered albedo, that is the existence of a snow cover. The beginning of year 2009 was exceptionally cold and a snow cover sustained long. To the contrary, the beginning of 2008 was exceptionally warm and snow was present only sporadically.

The downward directed long-wave radiation reflects also the occurrence of clouds and could be an indicator of the heat island effect; \(LW \downarrow\) is often enhanced in urban areas by the warmer than normal surrounding air masses, as mentioned above. The 5-day running mean varies approximately between 370 (summer) and 190 \(\text{ Wm}^{-2}\) (winter) and has a synoptic scale variation amplitude of about \(100 \text{ Wm}^{-2}\). The seasonal variation stems primarily from the variation of surface temperature. For instance, the average daily value for February 2007, a cold month, is \(240 \text{ Wm}^{-2}\) and for February 2008, a warm month, \(285 \text{ Wm}^{-2}\). The amplitude of the diurnal variation is modest and not as important a factor as the long-term variation of the surface temperature (not shown). The upward directed long-wave radiation is a direct measure of the surface temperature (equation \(2.55\)). The seasonal variation reaches from 230 to \(440 \text{ Wm}^{-2}\) corresponding to a range of \(-19\) to \(25^\circ\text{C}\) (assuming an emissivity of 0.98). Again, the temperature difference between Februaries 2007 and 2008 is evident in the \(LW \downarrow\) time series.

The net radiation is formed from the four components \((SW \downarrow, SW \uparrow, LW \downarrow, LW \uparrow)\), as mentioned above. Its 5-day running mean is slightly negative during winter and reaches about \(190 \text{ Wm}^{-2}\) in summer. Hour-to-hour data vary between -80 and \(620 \text{ Wm}^{-2}\). The diurnal variation is always as important as the seasonal and synoptic scale variation (not shown). The average diurnal course of \(R_n\) shows negative values at nights at all times of a year indicating long-wave cooling. In December and January \(R_n\) hardly gets positive values at all on average.

The comparison of radiation components measured at different sites is highly problematic. First of all, the latitude and altitude affect the short-wave radiation
Figure 4.11: Time series of the net radiation ($R_n$) and its four components (equation 2.55). Gray dots are half an hour values and black lines 5-day running means.
input. Second, the albedo differs between sites and is often not well reported. Third, measurements at some sites are actually done inside canopy and are thus affected by a reduced sky factor. Fourth, the extent of the heat island effect has an influence on $LW_\downarrow$.

### 4.3.2 Sensible and latent heat fluxes

The time series of $H$ represented in the following are corrected using the theoretical transfer function correction. The $LE$ values are corrected using the experimental transfer function correction having $\tau_s$ as a constant.

As stated before, in urban environments the restricted availability of water causes the channeling of available energy to sensible heat. SMEAR III is an exception in the sense that water availability is unrestrained in the vegetation sector. $LE$ is on average about 35 Wm$^{-2}$ in the urban sector during summer whereas the corresponding values for the road and vegetation sector are 60 and 80 Wm$^{-2}$ (not shown). $H$ behaves oppositely as it gets the smallest values in the vegetation sector. The time series of $H$ and $LE$ (figure 4.12) show slightly larger fluxes of sensible heat. Both fluxes exhibit a clear seasonal pattern having a maximum in summer and a minimum in winter. The monthly average diurnal courses of $H$ show negative values during nights and positive values during day time for most of the months (not shown). Interestingly, in winter, $H$ is positive during night indicating an excess source of energy which stems from either $Q_A$, $Q_S$ or $Q_F$. The diel course of $LE$ is in the same phase as $H$ but negative values are not observed. For diurnal courses of $H$ and $LE$ at SMEAR III, see Vesala et al. (2008). They made an analysis of $H$ and $LE$ in addition to the Bowen ratio ($\beta = H/LE$) and show the diurnal courses of these three variables for all of the wind sectors and for winter, spring and summer. Their data record extends from December 2005 to August 2006.

The energy flux values can be compared to measurements at SMEAR II, a forest site in Middle-Finland. Suni et al. (2003) show extensive time series of $H$ and $LE$ from 1996 to 2000. The sensible heat flux is generally measured to be smaller over the forest than at the urban site, as can be expected. More negative values of $H$ are observed at SMEAR II especially in winter. The latent heat flux, on the other hand, is higher at the forest site. Suni et al. (2003) also show the average diurnal course of the turbulent energy fluxes in August 2001. The maximum $H$
is around 150 Wm$^{-2}$ at SMEAR III whereas the corresponding value in Helsinki ranges from 120 to 280 Wm$^{-2}$, depending on the year. The maximum values for $LE$ are 200 Wm$^{-2}$ and 120 to 200 Wm$^{-2}$, respectively. Clearly, the energy portioning differs between the sites as was expected from the reasoning in section 2.5.3.

4.3.3 Residual term

The time series of the residual of the energy balance using an open- and a closed-path analyzer are also shown in figure 4.12. The 5-day running means are negative during winter indicating that the energy in the turbulent fluxes is not supplied by $R_n$. During summer, $Res$ is positive indicating heat release from the ground (assuming $Q_A = 0$) as $Q_F$ is always positive by definition. Figure 4.13 displays the monthly average diurnal course of $Res$ and gives more insight into the variation of the term. The open-path $LE$ is used in the figure since the data span is much larger. Generally the diurnal courses show negative values during nights and winter days. The average residual flux can be as strong as -100 Wm$^{-2}$ on winter nights whereas maximums are around 150 Wm$^{-2}$ on summer days. The variation amplitude is the largest for summer and smallest for winter. Cold temperatures (figure 4.1) almost always shift the pattern towards negative values as $Q_S$ is more negative and $Q_F$ more positive.

The three terms in $Res$ ($Q_A, Q_S, Q_F$) are not distinguishable, which is unfortunate since information on $Q_F$ would be especially interesting. Anyhow, the magnitude of the anthropogenic heat flux can be assessed in three rather speculative ways. These methods include (I) the shape of the diurnal course of $Res$, (II) comparison of $Res$ between land usage sectors and (III) the fact that $Q_S$ should sum to zero over a year.
Figure 4.12: Time series of sensible heat flux ($H$), latent heat flux ($LE$) and the residual of the energy balance. OP – open-path gas analyzer, CP – closed-path gas analyzer. Gray dots are half an hour values and black lines 5-day running means.
Figure 4.13: Diurnal variation of the residual in the energy balance. The calculations are done using the open-path $LE$ due to better data coverage.
First, something can be said about $Q_F$ using the shape of the diurnal courses of $Res$. $Q_S$ alone would have a quite symmetrical pattern peaking around noon (e.g. Offerle et al. (2005)) whereas the pattern of $Q_F$ has two peaks around 9:00 and 19:00 and a clear minimum around 5:00 local time (e.g. Sailor and Lu (2004)). According to this reasoning, $Q_F$ could be seen as more negative values in late evening compared to early morning. This is indeed the case for nights in late fall and early spring. The difference seems to be about 20 Wm$^{-2}$. The diurnal courses in winter have such a high degree of scatter that the pattern cannot be seen.

Second, $Res$ as a function of wind direction can also shed some light on the role of $Q_F$. Figure 4.14 displays a polar plot of $Res$ for all four seasons. The residual is again calculated using the open-path $LE$. In spring, the only negative values are observed in the urban sector which implies either the presence of a larger anthropogenic heat source or a larger heat release from a storage. This sector indeed contains building materials which have a higher capability of storing heat compared to for instance the vegetation sector. In summer, some negative values are also observed in the urban sector and the vegetation sector shows large positive values implying heat storage to the ground. The winter time figure seems most interesting. The residual is negative in all wind directions and the largest values are again in the urban sector, -45 Wm$^{-2}$, whereas the corresponding values for the road and vegetation sector are -30 and -25 Wm$^{-2}$. $Q_S$ can be argued to be small during winter in the urban sector, and if $Q_A$ is again neglected, $Q_F$ can be estimated to be 45 Wm$^{-2}$. It must be noted again, that energy fluxes at times when the wind is blowing from the urban sector might register microclimate anomalies as the roughness sublayer might exceed the measurement height.

Third, the sum of $Q_S$ should equal zero over a year as the ground is neither warming up nor cooling in the long term. Figure 4.15 displays the monthly average $Res$ for 2007-2009. Values are given for $Res$ calculated using the closed-path $LE$ without spectral corrections, with transfer function corrections without a relative humidity dependency and with corrections where $\tau(RH)$. The values are means of monthly average diurnal courses since a continuous record of $Res$ is not attainable with the current measurement systems. Year 2008 is the only year with full data coverage and the annual sum of $Res$ is -15.3, -65.3 and -59.0 Wm$^{-2}$ for uncorrected, $\tau=$constant and $\tau(RH)$ data. Consequently, $Q_F$ can be estimated to be about 60 Wm$^{-2}$, which agrees with the robust estimates above.
The choice of high frequency spectral correction has a large impact on the value of the residual, as seen from the annual sums above. Because spectral corrections increase $LE$, $Res$ should become more negative. The corrected values in figure 4.15 are indeed smaller than uncorrected values, and the inclusion of relative humidity dependency increases $Res$ slightly compared to a correction without $RH$ dependency. The percentual influence on annual averages is drastic. The correction without taking $RH$ into account changes the annual mean $Res$ by 35, 327 and 660%.

Figure 4.14: Residual of the energy balance in different seasons and as a function of wind direction. Blue dots denote negative values and red positive values. Gray lines are the 25 and 75 percentiles. The data is for unstable situations which ensures that the flux originates from relatively close to the sensors.

Figure 4.15: Average monthly residual in 2007-2009 without spectral corrections and with a spectral correction where $\tau(RH)$ and $\tau=\text{constant}$. Averages are taken from average diurnal courses of different months.
(4.52, 4.17 and 3.92 Wm\(^{-2}\)) for 2007-2009, whereas the corresponding values for a 
\(\tau(RH)\) correction are 36, 285 and 454\% (4.64, 3.64 and 2.70 Wm\(^{-2}\)).

Lately, \(Res\) has drawn more and more attention as a research subject as cities 
are growing and the climate is gradually warming. The additional heat release of \(Q_F\) 
and high \(Q_S\) increases urban temperatures and this might affect the health of city 
dwellers. Unfortunately, methodological imperfection is always present in estimating 
these terms. \(Q_F\) and \(Q_S\) are often treated together as a residual of the energy 
balance (neglecting \(Q_A\)), as in this study. Another approach is solve either one and 
then treat the other one as a residual. \(Q_F\) has often been calculated from energy 
consumption estimates and traffic flow rates, whereas \(Q_S\) has been solved using 
an objective hysteresis model where \(Q_S\) is a function of \(R_n\) and surface properties 
of the site (e.g. Roberts et al. (2006)). Also an Element Surface Temperature 
Method where a limited amount of surface temperature measurements are used and 
the heat transfer is modeled (Offerle et al. 2005). In addition, a numerical model 
(Town Energy Balance) to compute all the components in the energy balance has 
been recently developed (Masson et al., 2002). It is intended to be coupled with 
numerical weather prediction models.

Tables 4.2 and 4.3 show a collection of studies on \(Q_F\) and \(Q_S\). Note that in some 
studies a single value containing both terms is reported and is here shown in table 
4.3. The corresponding values of \(Res\) for SMEAR III are in table 4.3 though the 
definitions do not always match. The anthropogenic heat flux ranges from a couple 
of Wm\(^{-2}\) to even hundreds in Tokyo (Ichinose et al. 1999). Oke (1988) (not included 
in the table) also cited a set of energy consumption studies that give annual average 
values of 20 to 160 Wm\(^{-2}\). Most of the studies represented are conducted in cities 
at lower latitudes than Helsinki and thus the energy consumption due to heating is 
mostly lower, though ventilation might be as important a factor. Reykjavik (64\(^\circ\)N) 
seems to be the closest in latitude and a value of 35 Wm\(^{-2}\) was observed there 
(Steinecke, 1999). The values in table 4.2 are generally in consistence with the 
above estimates for SMEAR III.

Three studies (Oke et al. (1999), Grimmond and Oke (1999b), Lemonsu et al. 
(2004)) in table 4.3 give values for \(Q_S\) corresponding to the definition of \(Res\) at 
SMEAR III. Values differ of course due to different surface and climatic properties 
but clearly the orders of magnitude are the same. In cities at lower latitudes the 
amplitude of diurnal variation is larger than in Helsinki. When more heat is stored
during the day, also more heat is released during the night. All in all, also $Q_S$ seems to have daily mean values of a couple of tens of Wm$^{-2}$.

Defining $Res$ (or $Q_F$ and $Q_S$) is prone to several errors. First, it is difficult to ensure that the measurements are done in the ISL as the RSL is high in an urban environment. Second, the source areas of different components never coincide. The signal for $R_n$ comes from a steady circular area under the sensor whereas the source area (footprint) of the turbulent fluxes varies with stability and wind speed, to name but a few (section 2.2.3). All the measurement errors in $H$, $LE$, $R_n$ will accumulate in $Res$. Defining $Q_S$ directly from measurements is highly challenging because of the 3-dimensional structure of the urban surface and myriad material types used. Calculating $Q_F$ from energy consumption statistics is also a very robust approximation. Recent work also shows that advection plays an important role in the energy balance. Pigeon et al. (2007) calculated the advection of heat and moisture for Marseille, France, from one month summer data using conservation equations. The results show energy fluxes of 100 Wm$^{-2}$ and -50 Wm$^{-2}$ for $H$ and $LE$. Advection is most probably an issue in Helsinki too as it is a coastal city as Marseille. The current measurements conducted at SMEAR III unfortunately do not allow the investigation of the role of advection.
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<td>Christen and Vogt (2004)</td>
<td>Basel, Switzerland</td>
<td>2001-03</td>
<td>summing</td>
<td>Monthly mean range 7.0 W m$^{-2}$ and 10 W m$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>Grimmond et al. (2004)</td>
<td>Marseille, France</td>
<td>Jun-Jul 05</td>
<td>Estimated from other studies and CO$_2$ emissions</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Offerle et al. (2005)</td>
<td>Lodz, Poland</td>
<td>2001-02</td>
<td>ESTM</td>
<td>Jan-Aug -3 W m$^{-2}$, Oct-Mar 32 W m$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>Pigeon et al. (2008)</td>
<td>Basle, Switzerland</td>
<td>2000</td>
<td>ESTM</td>
<td></td>
<td>** 2008</td>
</tr>
</tbody>
</table>

Note that the negative value from Offerle et al. (2005) is non-physical. ** Energy consumption is used for deriving the components of Q$_S$ and Q. * The method, summing means that an inventory of energy consumption is used for deriving the components of Q. Table 4.3: Literature values for the anthropogenic heat flux. The method, summing means that an inventory of energy consumption is used for deriving the components of Q. * The method, summing means that an inventory of energy consumption is used for deriving the components of Q.
Table 4.3: Literature values for the storage heat flux. Method names: Res - a residual of the energy balance \((Q_F)\) incl. means that the term is included in \(Q_S\) whereas eval. means that \(Q_F\) is evaluated, OHM - objective hysteresis model, TEB - town energy balance model, TMS - thermal mass scheme, ESTM - element surface temperature method. See references for further information. ⋆ study also in table 4.2. ⋆⋆ compare with Grimmond et al. (2004).

The day time values for Helsinki are from 11-15 and night time from 22-04 o’clock.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Place</th>
<th>Data</th>
<th>Method</th>
<th>(Q_S) Day max</th>
<th>(Q_S) Night min</th>
<th>(Q_S) Ave</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oke et al. (1999)</td>
<td>Mexico City, Mexico</td>
<td>1 week</td>
<td>Res, (Q_F) incl. Dec93</td>
<td>280 Wm(^{-2})</td>
<td>-38 Wm(^{-2})</td>
<td>130 Wm(^{-2})</td>
</tr>
<tr>
<td>Grimmond and Oke (1999b)</td>
<td>Mexico City, Mexico</td>
<td>Dec93</td>
<td>Res, (Q_F) incl.</td>
<td>daily mean (-6 Wm(^{-2})</td>
<td>(-38 Wm(^{-2})</td>
<td>-1 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Vancouver, USA</td>
<td>Aug92</td>
<td>TEB</td>
<td>night min (-110)</td>
<td>day max (260 Wm(^{-2})</td>
<td>-38 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Chicago, USA</td>
<td>Jun-Aug95</td>
<td>TEB</td>
<td>night min (-18)</td>
<td>day max (24 Wm(^{-2})</td>
<td>-70,110 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Miami, USA</td>
<td>May-Jun95</td>
<td>TEB</td>
<td>night min (-105)</td>
<td>day max (150 Wm(^{-2})</td>
<td>-70,110 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Tucson, USA</td>
<td>Jun90</td>
<td>TEB</td>
<td>night min (-160)</td>
<td>day max (210 Wm(^{-2})</td>
<td>-70, 150 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>San Gabriel, LA, USA</td>
<td>Jul94</td>
<td>TEB</td>
<td>night min (-170)</td>
<td>day max (64 &amp; 80 Wm(^{-2})</td>
<td>-5, -18 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Vancouver, USA</td>
<td>Jul-Sep92</td>
<td>TEB</td>
<td>night min (-196)</td>
<td>day max (163 &amp; 184)</td>
<td>-8 &amp; -38 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Sacramento, USA</td>
<td>Aug91</td>
<td>TEB</td>
<td>night min (-18)</td>
<td>day max (39 Wm(^{-2})</td>
<td>-5 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Arcadia, LA, USA</td>
<td>Jul94</td>
<td>TEB</td>
<td>night min (-170)</td>
<td>day max (33 Wm(^{-2})</td>
<td>-10 Wm(^{-2})</td>
</tr>
<tr>
<td>Masson et al. (2002)</td>
<td>Mexico City, Mexico</td>
<td>1 week</td>
<td>TEB</td>
<td>night min (-110)</td>
<td>day max (260 Wm(^{-2})</td>
<td>-38 Wm(^{-2})</td>
</tr>
<tr>
<td></td>
<td>Vancouver</td>
<td>15d Aug92</td>
<td>TEB</td>
<td>night min (-105)</td>
<td>day max (150 Wm(^{-2})</td>
<td>-70,110 Wm(^{-2})</td>
</tr>
<tr>
<td>Lemonsu et al. (2004)</td>
<td>Marseille, France</td>
<td>Jun-Jul01</td>
<td>TEB</td>
<td>night min (-160)</td>
<td>day max (24 Wm(^{-2})</td>
<td>-70, 150 Wm(^{-2})</td>
</tr>
<tr>
<td>Christen and Vogt (2004)</td>
<td>Basel, Switzerland, 2 sites</td>
<td>2001-03</td>
<td>Res</td>
<td>day 163 &amp; 184, night 64 &amp; 80 Wm(^{-2})</td>
<td>-5, -18 Wm(^{-2})</td>
<td></td>
</tr>
<tr>
<td>Offerle et al. (2005)</td>
<td>Lodz, Poland</td>
<td>2001-02</td>
<td>ESTM</td>
<td>annual 0 Wm(^{-2})</td>
<td>Apr 6, Dec - 8</td>
<td>-38 Wm(^{-2})</td>
</tr>
<tr>
<td>Roberts et al. (2006)</td>
<td>Marseille, France</td>
<td>Jul 8d</td>
<td>Res, (Q_F) eval.</td>
<td>day 67 Wm(^{-2})</td>
<td>night -99 Wm(^{-2})</td>
<td>50, -7 Wm(^{-2})</td>
</tr>
</tbody>
</table>
Summary and conclusions

A three and a half year long micrometeorological eddy covariance data record was used to investigate energy fluxes in Helsinki, Finland. The first aim was to study the temporal variation of the energy fluxes having the residual of the energy balance as the main target. In addition, the effect of high frequency spectral corrections on water vapor fluxes measured with a closed-path IRGA was assessed. The study methods consisted of widely accepted eddy covariance data processing procedures (Aubinet et al., 2000), and numerical methods such as Fast Fourier transfer and Wavelet analysis were used.

The experimental high frequency spectral correction using the transfer function method corroborate the results of Ibrom et al. (2007), Massman and Ibrom (2008) and Mammarella et al. (2009): the response time of water vapor clearly depends on relative humidity. It is also evident that the transfer function cannot be determined theoretically for an active scalar such as water vapor, which is congruent with the results of Aubinet et al. (2000). A new wavelet-based spectral correction method that accounts for the variation attenuation by an amplitude adjustment was developed. The amplitude adjustment for H$_2$O led to desired results in raw data manipulation, but the correction of the measured covariances remained small. To the contrary, the transfer function based corrections with and without a RH dependency gave on average an 11% and a 14% addition to the measured flux. The reason for the discrepancy between the transfer function and wavelet-based corrections may be that the wavelet-based correction acts only to increase the covariance in the inertial subrange whereas the transfer function correction corrects the covariance at a wider range of frequencies. Perhaps also the answer lies in the assumptions of local and
global spectral similarity that were made in the transfer function method.

Extensive time series of net radiation $R_n$, sensible heat flux ($H$) and water vapor $LE$ were displayed. The differences between land use sectors was evident and available energy is clearly channeled to $H$ when $LE$ is restricted by water availability. The residual term ($Res$) was negative during winter and summer nights indicating an excess source of energy fueling $H$ and $LE$. The energy originates either from the heat release from the ground ($Q_S < 0$) or from the anthropogenic heat flux ($Q_F$). The magnitude of $Q_F$ was estimated in three approximative ways: (I) asymmetry in the diel course of $Q_S$, (II) $Q_S$ can be assumed to be small in the urban sector during winter (figure 4.14) and (III) the sum of $Q_S$ over a year should equal zero (figure 4.15). The methods gave values of about 20 Wm$^{-2}$ for winter nights, 50 Wm$^{-2}$ for winters and 60 Wm$^{-2}$ for the whole year. The results agree with the findings represented in a wide literature review (table 4.2).

The role of spectral corrections in an urban environment as an opposite to simpler terrains is pronounced. The reason is the growing interest in the anthropogenic heat flux and the heat storage to the building-air volume. These two terms are solved often as a residual of the energy balance, as in this study, and hence any small alterations in the energy fluxes have a large effect on the residual. The spectral corrections are also more demanding in a heterogeneous environment like a city. The high degree of heterogeneity can for instance lead to spectral dissimilarity between scalars. The effect of the relative humidity dependent spectral correction on the annual sum of $Res$ was drastic: a difference of 36 to 454%, corresponding to 4.64 and 2.70 Wm$^{-2}$, was observed depending on the year.

In conclusion, it can be stated that the research on urban environments is becoming more and more important due to the ever-growing fraction of urban citizens. The inadvertent climate modification resulting from city growth stems ultimately from surface-atmosphere interaction. The flux of anthropogenic heat and the heat storage to a building-air volume draw most attention as they cannot be accurately measured and are the main reasons for climate alteration. Furthermore, the present way of solving the terms as a residual of the energy balance causes the problem of accurate energy flux measurements. Consequently, all small corrections of the energy fluxes in urban areas must be done with a high degree of precision.


Wilson, K., A. Goldstein, A. Falge, M. Aubinet, D. Baldocchi, P. Berbigier, C. Bernhofer, R. Ceulemans, H. Dolman, C. Field, A. Grelle, A. Ibrom, B. E. Law,