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EXTENDING THE APPLICABILITY OF THE
EDDY-COVARIANCE FLUX-MEASUREMENT
TECHNIQUE

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Academic dissertation

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Annika Iida Nordbo
University of Helsinki, 2012

Abstract

Surface–atmosphere exchange of momentum, energy and atmospheric constituents affects the atmosphere—from alterations in local microclimates and mesoscale weather to climate modification. These exchange processes can be studied using direct eddy-covariance (EC) measurements of vertical turbulent transport, but the technique has not yet readily been applied in non-prevailing ecosystems. Thus, the aim of this thesis is to extend the applicability of the EC technique in two ways: to non-standard sites and by further developing the technique itself. To reach the aim, EC measurements over a boreal lake and three urban sites in Helsinki were performed.

Long-term measurements over a lake revealed that the water below the thermocline was decoupled from the atmosphere and thus not important for atmospheric vertical turbulent fluxes. The energy exchange between the lake and the atmosphere departs from vegetated surfaces especially due to large nocturnal evaporation fuelled by lake-water heat storage. Long-term measurements at a semi-urban site in Helsinki showed that the surface–atmosphere exchange is altered by anthropogenic activity: changes in surface-cover and an additional anthropogenic heat release (13 W m^{-2}) led to an altered surface energy balance, and anthropogenic CO_2 emissions led to a large positive annual CO_2 balance (1.8 kg C m^{-2}). Intra-site and intra-city variation in surface-cover led to differences in atmospheric stability and CO_2 emissions. The EC technique evaluation demonstrated that (i) the 'energy imbalance problem' in EC measurements is not primarily surface-cover dependent, and that (ii) common calculation errors in EC calculations can be almost 30% of the flux. Water vapour flux measurements with a closed-path analyser were affected by sorption: the signal's arrival is delayed and it is attenuated. A new spectral-correction method based on wavelet analysis was developed to automatically correct for this signal attenuation of constituents.

The conclusions of this thesis improve the understanding of surface–atmosphere exchange over non-standard ecosystems. The lake measurements will continue to be used for improving weather forecasts, and the results from the urban studies can be used in city-planning. The EC technique is developed by offering guidance in calculations at urban sites and by introducing a new correction algorithm.

Keywords: eddy covariance, turbulent flux, surface energy balance, CO_2 flux, spectral correction, wavelet analysis

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List of publications

This thesis consists of an introductory review, followed by 5 research articles. The papers are reproduced with the permission of the journals concerned. In the introductory part, these papers are cited according to their roman numerals.

- I **Nordbo A.**, S. Launiainen, I. Mammarella, M. Leppäranta, J. Huotari, A. Ojala and T. Vesala (2011). Long-term energy flux measurements and energy balance over a small boreal lake using eddy covariance technique. *J. Geophys. Res.*, **116**, D02119, doi:10.1029/2010JD014542.
- II **Nordbo A.**, L. Järvi and T. Vesala (2012). Revised eddy covariance flux methodologies – effect on urban energy balance. *Tellus B*, **64**, 18184, doi:10.3402/tellusb.v64i0.18184
- III Järvi L., **A. Nordbo**, H. Junninen, A. Riikonen, J. Moilanen and T. Vesala (2012). Seasonal and annual variation of carbon dioxide exchange in Helsinki, Finland, in 2006-2010. *Atmos. Chem. Phys.*, **12**, 8475-8489, doi:10.5194/acp-12-8475-2012
- IV **Nordbo A.**, L. Järvi, S. Haapanala, J. Moilanen and T. Vesala (2012). Intra-city variation in urban morphology and turbulence structure in Helsinki, Finland. *Bound.-Layer Meteorol.*, doi:10.1007/s10546-012-9773-y
- V **Nordbo A.** and G. Katul (2012). A wavelet-based correction method for eddy-covariance high-frequency losses in scalar concentration measurements. *Bound.-Layer Meteorol.*, doi:10.1007/s10546-012-9759-9

1 Introduction

At the end of 2011, 52% of the World’s population was estimated to live in cities ([United Nations Population division, 2012](#)) and estimates of the extent of urbanization vary from 0.2% to 2.4%, depending on the definition and means of data collection ([Schneider et al., 2009](#)). In Finland, urban areas cover 4% of the land surface (Figure 1). The high population density and alteration in surface cover have led to inadvertent climate modification primarily at the local scale. The development of an urban microclimate is caused by the anthropogenic changes in surface–atmosphere exchange: the vertical turbulent transport of momentum is increased, energy partitioning to evaporation and direct heat is altered and there is an excess of CO₂ and heat release in cities ([Roth, 2000](#); [Arnfield, 2003](#); [Velasco and Roth, 2010](#); **Papers II, III, IV**). These modifications affect human comfort—e.g. through air quality and the urban heat island effect ([Oke, 1982](#))—and weather forecasts through alterations in the surface energy balance. Furthermore, over 70% of global energy-related carbon dioxide (CO₂) emissions are estimated to originate from cities ([WEO, 2008](#)), which makes urban areas the hotspots of potential reductions in greenhouse-gas emissions. These urban microclimate alterations have together risen the need for direct measurements and the need to increase the understanding of the vertical exchange processes between an urban surface and the atmosphere. Urban areas also have mesoscale effect through pollution and excess heat.

Globally, lakes cover over 3% of the continental surface ([Downing et al., 2006](#)), but the fraction is higher—approximately 7%—in the boreal zone a distinct band of coniferous trees which circles the entire northern hemisphere. In Finland the fraction of lakes is 10% (Figure 1) but may exceed even 35% in some areas (Figure 2). These lakes are vital in terms of freshwater resources and surface–atmosphere interaction via physical processes. Lakes can also be used as sentinels of climate change, since they are very sensitive to global warming ([Williamson et al., 2009](#)). Like urban areas, lakes alter the surface–atmosphere exchange of energy, and the high thermal capacity of water causes the air temperature of lake regions to lag those of the surrounding land areas. Despite their large area coverage and distinct energy exchange, research on lakes is still scarce compared with vegetative surfaces (**Paper I**). Lakes have also been parametrized crudely in numerical weather prediction, although they are known to affect the local climate ([Bonan, 1995](#); [Long et al., 2007](#); [Dutra et al., 2010](#); [Samuelsson et al., 2010](#); [Balsamo et al., 2012](#)). A major reason for the exclusion has been the lack of a lake coverage and depth dataset ([Kourzeneva et al., 2012](#)). Lakes are often not taken into

account at all or are taken to have properties of nearby oceans. Recently, this issue has been improved by including separate lake modelling parametrizations in large-scale models (Dutra et al., 2010; Eerola et al., 2010; Mironov et al., 2010; Martynov et al., 2012), which further emphasizes the need for measuring the vertical exchange between lakes and the atmosphere.

The surface–atmosphere exchange over urban areas and lakes can be studied by applying the eddy-covariance (EC) flux-measurement technique (for a review see e.g. Aubinet et al., 2012). It is the only direct means to measure vertical turbulent transport of momentum, heat and constituents (e.g. CO₂ and aerosol particles). It provides near-automatic, long-term measurements typically with a 30-minute resolution. The good short-term and long-term representativeness of the measurements enables the study of ecosystem dynamics and the calculation of annual budgets of greenhouse gases. The EC technique has been accepted as the conventional means of studying surface–atmosphere exchange and it has been widely applied in vegetative environments (Valentini et al., 2000; Baldocchi et al., 2001; Mizoguchi et al., 2009).

This thesis work addresses some of the challenges and needs for better understanding of the lake–atmosphere and urban–atmosphere exchange of properties. The **generic aim** of this work is to broaden the applicability of the eddy-covariance technique (i) to less-studied and complex environments, such as lakes and urban surroundings, and (ii) by revising and developing new flux-calculation methods. The **elaborated aims** of this work are to:

1. increase our knowledge of the thermal structure of lakes, quantify their surface–atmosphere energy exchange and determine the energy balance closure of a particular boreal lake;
2. determine the seasonal, inter-annual, and intra-city variation in urban turbulent surface–atmosphere exchange;
3. examine the annual CO₂ budget over various urban land-cover types and over several years;
4. quantify the effect of flux calculation procedures on urban energy balance closure and evaluate the systematic errors and random uncertainties in urban EC measurements;
5. develop and verify a new method for correcting for the loss of small-scale turbulent EC fluxes.

Figure 1: Surface cover fractions in Finland. Water includes lakes and rivers. Data are from the Finnish Environment Institute (www.environment.fi).

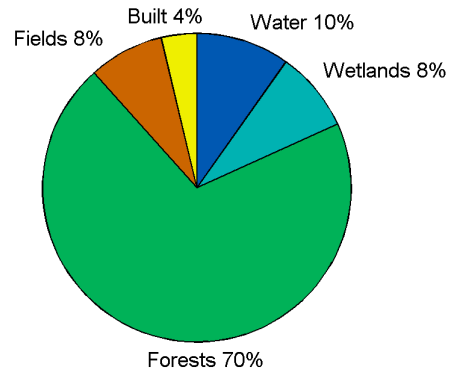
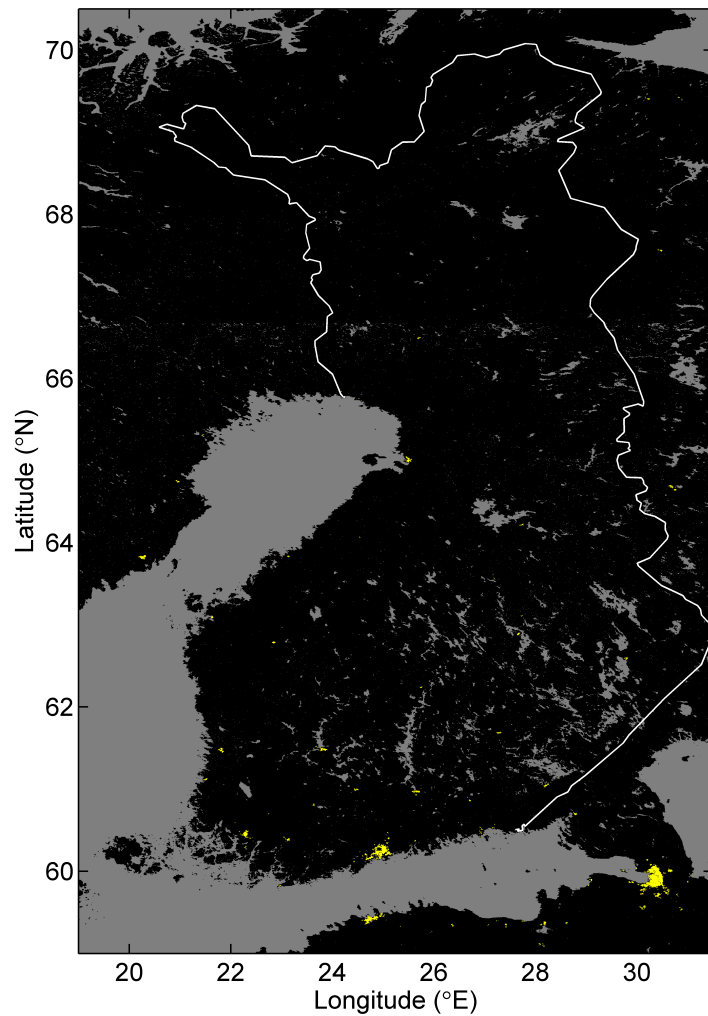


Figure 2: Surface cover types in Finland: non-urban land (black), urban land (yellow) and water (grey). The map is made using 500-m-resolution data from the MODIS satellite (Schneider et al., 2009, 2010). Only the country border of Finland is shown.



2 Background

2.1 The atmospheric boundary layer

The atmospheric boundary layer (ABL) is the lowest part of the atmosphere and is up to tens or thousands of metres deep, depending on atmospheric stability and the underlying surface. The definition of the ABL height is not unambiguous, and the layer is often considered as the volume of air that is affected by changes on the surface within about an hour of their occurrence (Stull, 1988). The ABL height—or the mixing layer height—can be determined for instance from the gradient of aerosol backscatter or from the variance of vertical wind (Eresmaa et al., 2006; Barlow et al., 2011a). The air in the ABL is always turbulent—containing random motion of 3-dimensional eddies—and vertical transport through turbulent mechanisms is always superior to that of molecular diffusion, except in the nearest millimetre to the ground. This turbulent transport results in net fluxes of momentum, heat and constituents (e.g. H₂O and CO₂).

The ABL can be divided into several vertical scales depending on the underlying surface type (Figure 3). Classically, the vertical structure of the ABL can be divided into an Ekman layer that overlies a surface layer near the ground. In the Ekman layer, fluxes decrease and winds veer with height. To the contrary, in the surface layer fluxes stay partly constant with height and the wind speed profile mainly follows a logarithmic behaviour in neutral conditions, modified by the Monin-Obukhov similarity (MOS) theory under stable or unstable conditions (orange in Figure 3). The surface layer in an urban environment, like over forests, can further be divided into a roughness sublayer (RSL, Raupach, 1979; Oke, 1987), where the flow consists of turbulence dominated by individual elements or sources and fluxes are not constant with height (Grimmond et al., 2004); and into an inertial sublayer (ISL), where fluxes are almost constant with height and traditional micrometeorological theories apply (Roth, 2000). An example of a traditional micrometeorological theory is MOS that describes the variation of mean flow and turbulence properties with height as a function of the Monin–Obukhov parameters (Obukhov length is described later). Consequently, EC measurements above a rough surface should always be conducted in the ISL, if one exists.

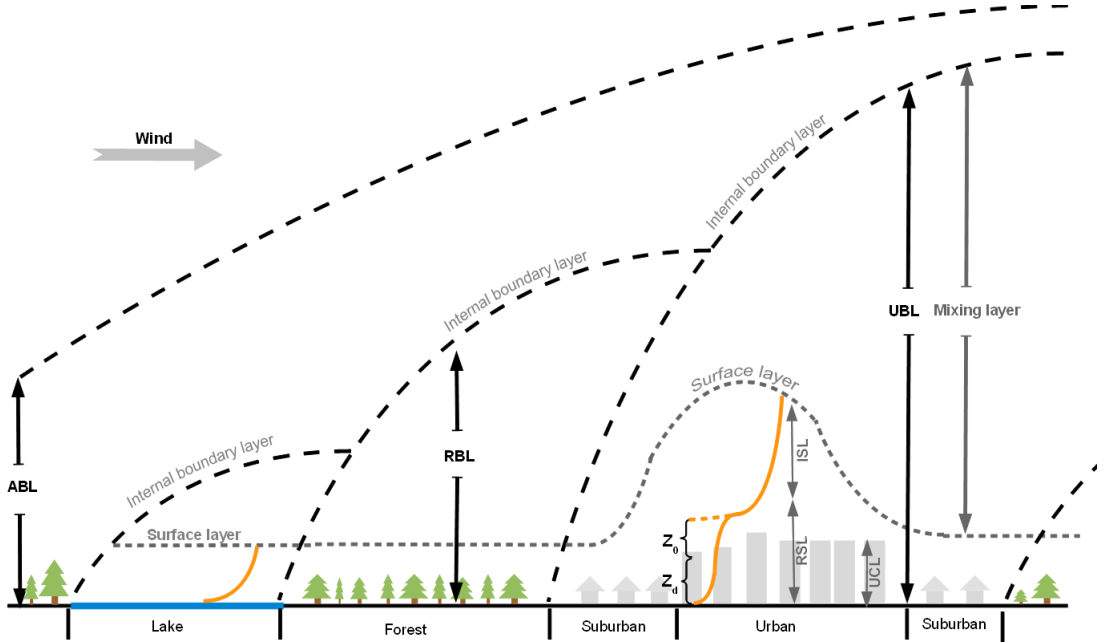


Figure 3: Schematic structure of the atmospheric boundary layer (ABL) that comprises internal boundary layers between dominant surface-cover types (lake, forest, suburban, urban; not to scale). RBL – rural boundary layer, UBL – urban boundary layer, RSL – roughness sublayer, ISL – inertial sublayer, UCL – urban canopy layer. Logarithmic wind profiles in the near-neutral surface layers (orange) are shown with the displacement height (z_d) and roughness length (z_0) over the urban surface. The urban boundary-layer structure is adapted partly from [Oke \(1987\)](#).

The wind profile in the ISL depends on the underlying surface. The profile over a lake follows micrometeorological theories starting already a few millimetres above the surface, but the urban profile reaches this state only near the top of the RSL (Figure 3; [Feddersen, 2005](#)). Over a lake, mean wind speed goes to zero below a height described by the aerodynamic roughness length ($z_0 \approx 1$ mm); over an urban canopy, the logarithmic wind profile in neutral atmospheric conditions would go to zero at a height $z_0 + z_d$, where z_d is the zero-plane displacement height, which describes the height where most of momentum is absorbed. The urban roughness parameters, z_0 and z_d , are theoretical concepts and they can be estimated from wind profile measurements or based on urban morphology ([Grimmond and Oke, 1999](#); **Paper IV**). The z_0 over a water surface can be parameterized with momentum transfer ([Smith, 1988](#)) and $z_d = 0$, by definition.

The sharp horizontal changes in surface cover, for instance between a lake and an urban area, result in horizontal internal boundary layers within which the airflow is adapting to the underlying surface (Figure 3, Garratt, 1990; Cheng and Castro, 2002; Barlow et al., 2008). These internal boundary layers are the borders of the layers belonging to different surfaces, and the development of the internal boundary layer depends on atmospheric stability and surface roughness.

2.2 Eddy-covariance technique

The eddy-covariance (EC) technique has been the conventional approach to measure vertical turbulent fluxes for about 20 years. The technique offers direct estimates for momentum, heat, and constituent exchange between the surface and the atmosphere provided the flow is turbulent, stationary, horizontally homogeneous and subsidence is not present (Baldocchi et al., 1988; Lee et al., 2004; Aubinet et al., 2012). The EC technique relies on measuring the simultaneous fluctuations in vertical and horizontal velocity components (w' , u' , v') and a scalar (s') of interest (temperature, H₂O, CO₂) in the inertial sublayer. The turbulent scalar flux is related to the covariance of the vertical wind and the scalar, and the covariance can be calculated as

$$\overline{w's'} = \frac{1}{N} \sum_{i=1}^N (w_i - \bar{w})(s_i - \bar{s}), \quad (1)$$

where w_i is the measured instantaneous vertical wind speed, \bar{w} is the mean vertical wind speed, s_i is the measured instantaneous scalar concentration, \bar{s} is the mean scalar concentration. N is the amount of data, and it is determined by the averaging time used for the covariance calculation, commonly taken to be 30–60 minutes. The averaging time depends on the size of eddies that are desired to be captured and thus depends on measurement height, overlying surface and atmospheric stability (Finnigan et al., 2003; Paper II). The friction velocity, which describes momentum transport, is calculated similarly as $u_* = \sqrt{|\overline{w'u'}| + |\overline{w'v'}|}$ (m s⁻¹), where u' is the horizontal wind fluctuation in the prevailing wind direction and v' is the horizontal wind fluctuation perpendicular to the prevailing wind direction.

The EC measurement setup consists traditionally of an ultrasonic anemometer, for measuring the three wind components and sonic temperature, and of an infrared

gas analyser (IRGA) for measuring concentrations of gases. The measurement frequency should be high, sometimes up to 20 Hz, in order to include the smallest energy-containing turbulence. This is especially important when measuring near a smooth surface, such as a lake (**Paper I**). Moreover, the IRGA is required to be located as near to the anemometer as possible, without causing too much flow distortion. There are two primary types of analysers: the open-path analyser measures the gas concentrations within the air next to the anemometer and the closed-path analyser is situated away from the anemometer and air samples are drawn through a tube to the analyser. The open-path IRGA suffers from problems related to flow distortion, surface heating and precipitation, and the closed-path IRGA suffers from a dampening of fluctuations along the tube’s length.

EC measurements are generally prone to several systematic errors due to the complexity of the setup, the state of the flow or the raw data post-processing (**Paper II**, Mauder et al., 2008). Flow distortion or vertical deflection caused by adjacent obstacles or the setup itself are especially common in an urban environment (Barlow et al., 2011b). Horizontal advection or storage beneath the measurement height may cause the measured flux to depart from the actual surface flux. Moreover, typical errors caused by insufficient raw-data processing are related to de-spiking, airflow coordinate rotations, de-trending, the determination of the lag time between the anemometer and a closed-path IRGA, surface heating problems of open-path gas analysers and spectral corrections (which account for low- and high-frequency loss of covariance). The corrections to avoid these errors are especially important in urban surroundings where the unmeasurable heat storage change and anthropogenic heat release are often estimated using measurements of other energy balance components (shown later).

The high-frequency spectral corrections have attracted more attention during the past decade, especially concerning H₂O flux measurements with a closed-path setup (Su et al., 2004; Ibrom et al., 2007a,b; Mammarella et al., 2009; Runkle et al., 2012; **Paper II**). So far these spectral corrections have been made in two primary ways: (i) using theoretical equations, including information on the measurement setup and wind speed and (ii) assuming that the sensible heat flux measurements (or \overline{wT}) are not attenuated and can thus be used as a perfect model in an experimental determination of the loss of other fluxes. The theoretical method suffers from the lack of background for taking into account all possible sources of attenuation, and the experimental method is laborious and suffers from an assumption of scalar similarity (i.e. heat, CO₂ and H₂O are assumed to be transported by similar eddies). Scalar similarity has recently

been shown with a co-spectral budget analysis to fail if the distribution of sources and sinks of the scalars concerned are not the same (Cava and Katul, 2012), though similarity applies in more heterogeneous sites (e.g. Runkle et al., 2012). Evaporation measurements with a closed-path analyser are especially complicated due to adsorption and desorption effects that water vapour undergoes when travelling down a sampling tube (Massman and Ibrom, 2008).

The source area (i.e. flux footprint) of the measurements, which can only in practice be estimated via modelling, may depart from the targeted area and cause an additional error (Göckede et al., 2008). The flux footprint can be described with a continuous contribution function that extends to zero at infinity. Commonly a certain contribution level, e.g. 80%, is chosen and calculated as a volume integral of the contribution function (Figure 4). The footprint area with a major contribution to the flux is often the size of a few hectares, depending on the measurement height, surface roughness and atmospheric stability (Vesala et al., 2008a). The footprint can be calculated analytically (e.g. Kormann and Meixner, 2001) or numerically (e.g. Sogachev and Lloyd, 2004; Markkanen et al., 2009). The footprint is mostly upwind of the tower, if measurements are conducted above the RSL, and thus it changes with wind direction. This can be seen as an advantage of the technique since one setup can measure several surface-cover types around a particular tower, but simultaneously this feature is disadvantageous when calculating annual balances of surface-atmosphere gas exchange (Paper III). Nevertheless, the surface should be fairly homogeneous in any given direction so that advection can be neglected and fluxes measured from a tower correspond to surface fluxes.

In addition to surface fluxes, other variables can be calculated from the EC measurements. The atmospheric stability (ζ) is a function of the sensible heat flux (H , W m^{-2}) and u_* , and can thus be directly estimated from EC data as

$$\zeta = \frac{z - z_d}{L}, \quad L = -\frac{\rho_a c_{pa} u_*^3}{\kappa \frac{g}{T} H}, \quad (2)$$

where z is measurement height (m), L is the Obukhov length (m, Obukhov, 1946), ρ_a air density (kg m^{-3}), c_{pa} air heat capacity at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$), $\kappa \approx 0.4$ is the von Kármán constant, g gravitational acceleration (m s^{-2}), T air temperature (K). The stability parameter is widely used in boundary-layer meteorology and in the MOS theory.

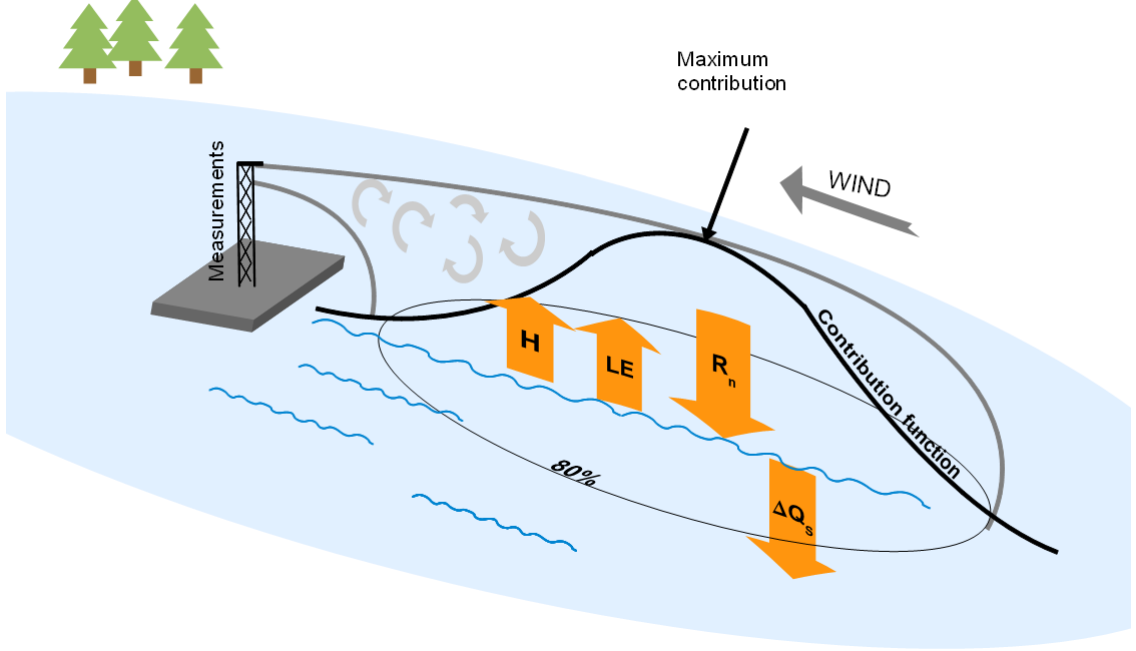


Figure 4: Schematic of the flux source area (footprint area) and the flux contribution function (footprint function) at a lake and the dominating components of the energy balance: H —sensible heat flux, LE —latent heat flux, R_n —net radiation, ΔQ_s —change in water heat storage. Orange arrows show the direction of a positive flux.

The wind profile in the surface layer also depends on u_* as

$$U(z) = \frac{u_*}{\kappa} \left[\ln \left(\frac{z - z_d}{z_0} \right) + \psi(z - z_d, z_0, L) \right], \quad (3)$$

where ψ is the stability term which goes to zero for neutral atmospheric stability ($\zeta \approx 0$). The term $(\frac{u_*}{U})^2$ is also known as the drag coefficient, C_D , which is used widely in wind profile estimation and dispersion modelling. Thus, the measurements of C_D can be used to evaluate the success of the determination of z_d and z_0 using morphological methods (**Paper IV**). Moreover, standard deviations of wind speed and scalars (or integral turbulence characteristics, [Monin and Yaglom, 1975](#)) can describe the features of the turbulent transport and the similarity between scalars (**Paper IV**). The power-spectra can reveal the sizes of eddies that are responsible for the variance of a certain variable, whereas co-spectra can reveal the sizes of eddies that cause the turbulent flux (**Papers IV, V**).

The power and co-spectra can be calculated using two alternative numerical methods: Fourier transforms and wavelet transforms (Farge, 1992; Mallat, 1999). In Fourier transforms, the time series is decomposed into periodically oscillating sine and cosine waves, whereas in wavelet analysis the data are decomposed into small pulses (a.k.a. wavelets). In both methods, the amplitude of the basis function (sine, cosine and wavelet) can be extracted for each frequency of oscillation. The amplitudes of variation as a function of frequency can then be used for calculating the power and co-spectra of the turbulent transfer processes (Stull, 1988). The advantage of wavelet analysis is that it not only preserves information on the frequency of oscillation but also the time when the oscillation happened. This so-called time–frequency localization was one of the original motivations behind the development of the method in the 80s (Mallat, 1999).

2.3 Surface–atmosphere interactions

2.3.1 Energy fluxes

All surfaces exchange energy with the atmosphere primarily through long-wave radiation and turbulent fluxes. The full **surface energy balance** contains the net all-wave radiation (R_n), surface heat storage change (ΔQ_S), turbulent sensible heat flux (H), turbulent latent heat flux (LE), the effect of precipitation (Q_P) and other site-specific terms (Q_I)

$$R_n - \Delta Q_S = H + LE + Q_P + Q_I. \quad (4)$$

Here H and LE are defined positive when directed upward, ΔQ_S is positive when the surface is heating and R_n is positive when the surface is gaining radiative energy.

In a full **lake surface energy balance**, ΔQ_S is the water heat storage change and Q_I includes the heat storage change in the bottom sediments and the effect of run-off. In turbid Finnish lakes, the role of bottom sediments can often be neglected and Q_P is generally of minimal importance, and the lake energy balance is described by only four terms (Figure 4, **Paper I**). The energy balance is dominated by two large terms, R_n and ΔQ_S , often together called the available energy. R_n is slightly larger than over land due to a lower water surface albedo (Beyrich et al., 2006); ΔQ_S is substantially larger than over land, and its dominance is enabled by the high thermal and heat transport

capacity of water. The large ΔQ_S also causes a drastic thermal lag in the lake system compared to surrounding land areas (Rouse et al., 2008). Evaporation also dominates over sensible heat, since it is not restricted by water-availability. Furthermore, the smooth lake surface causes a small momentum transport from the atmosphere to the surface, in comparison with vegetative surfaces. In summary, the relative importance of all the energy balance components and the momentum transport depend on the lake size, depth and location (Oswald and Rouse, 2004).

Two terms of the lake energy balance can be measured with the EC technique, namely H and LE . The EC technique is known to underestimate the sum of these fluxes in various ecosystems by about 20% compared with the available energy, $R_n - \Delta Q_S$ (Wilson et al., 2002; Foken et al., 2006). The main causes of this imbalance may be the system’s inability to detect turbulent transport by large eddies and the exclusion of transport by vertical advection. This energy imbalance is a major unresolved issue within the eddy-covariance community.

At lakes, long-term energy flux measurements using the EC technique have mainly concentrated on high-latitudes. The Great Slave Lake, Canada, is perhaps the most studied one (Blanken et al., 2000, 2003; Eaton et al., 2001; Rouse et al., 2003; Schertzer et al., 2003; Rouse et al., 2005, 2008) followed by research on other North-American lakes (Anderson et al., 1999; Eugster et al., 2003; Granger and Hedstrom, 2011) and a reservoir (Liu et al., 2009b, 2011). The energy exchange of inland water bodies have also been monitored in short-term measurements in Europe (Venäläinen et al., 1998; Panin et al., 2006; Jonsson et al., 2008; Vercauteren et al., 2008; Salgado and Moigne, 2010; Bouin et al., 2012), Israel (Assouline et al., 2008; Tanny et al., 2008, 2011) and Australia (McJannet et al., 2011). None of these studies provide a long-term energy balance closure analysis. Studies applying other flux estimation techniques (such as the gradient technique) also exist but these studies are not included here due to the indirect nature of the techniques.

In a full **urban surface energy balance**, Q' contains an important additional heat source, the anthropogenic heat flux Q_F (Sailor, 2011). The global mean urban Q_F has a diurnal range of 0.7–3.6 W m⁻², and is larger on weekdays compared with weekends (Allen et al., 2011). Furthermore, ΔQ_S is the heat storage in the building-air volume and the ground, and it is about an order of magnitude smaller than its lake counterpart. In an urban environment, ΔQ_S increases as a function of an active built index, and LE fraction increases as a function of an active vegetated index (Piringer and Joffre, 2005; Loridan and Grimmond, 2012). The effect of precipitation can again

be assumed negligible and the urban surface energy balance consists of five major components (Figure 5, **Paper II**). The surface, in this case, is not the ground but is taken as an imaginary surface slightly above the mean building height (z_H). Moreover, ΔQ_S and Q_F are hard to measure and are thus often taken together as the energy balance residual $Res = R_n - H - LE = \Delta Q_S - Q_F$, which is positive for a building-air volume losing heat ($\Delta Q_S < 0$) or if there is an anthropogenic heat release ($Q_F > 0$). ΔQ_S and Q_F can also be determined indirectly using modelling approaches (Roberts et al., 2006; Järvi et al., 2011), and some components of Q_F may be determined from pulses of high concentration in EC data (Kotthaus and Grimmond, 2012).

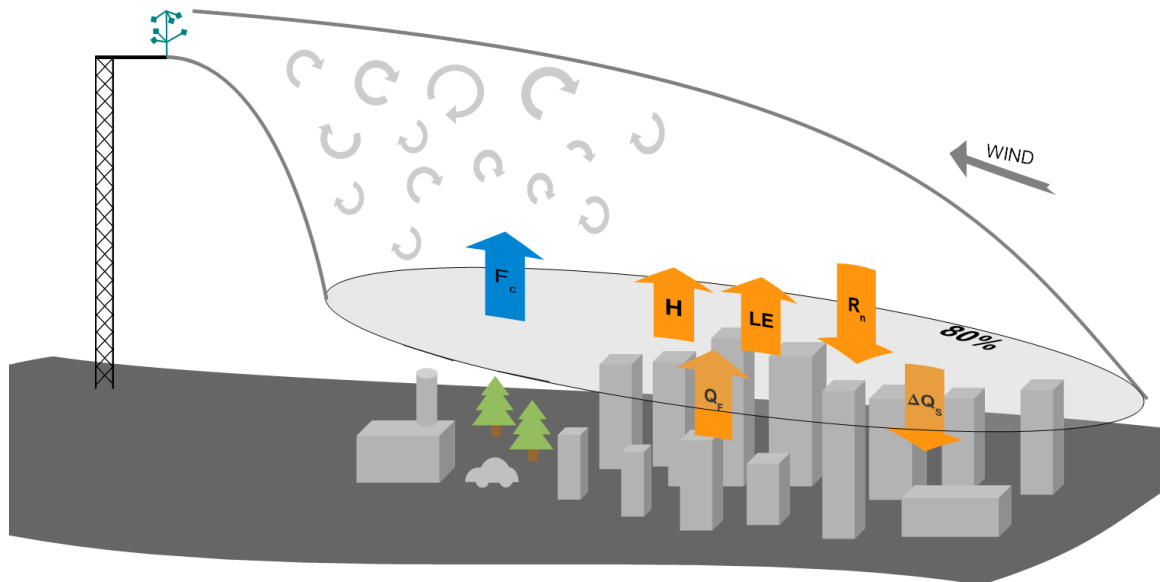


Figure 5: Schematic of eddy-covariance flux measurements in an urban surrounding. The flux source area (footprint area) is depicted with an oval, CO_2 flux (F_c) with blue arrow and the dominating components of the energy balance with orange arrows: H —sensible heat flux, LE —latent heat flux, Q_F —anthropogenic heat flux, R_n —net radiation, ΔQ_S —change in heat storage. Arrows show the direction of a positive flux and the situation would be typical during a summer’s day with CO_2 emissions.

The urban energy balance is prevalingly driven by the mainly rough and impervious surface: the momentum transfer (or u_*) is larger than over smoother surfaces, and the impervious surfaces restrict evaporation and lead to high sensible heat flux. The elevated H , the heat input from Q_F and the role of ΔQ_S together cause the urban

heat island effect, which is seen as elevated temperatures compared to surrounding rural areas, especially at night (Oke, 1982; Arnfield, 2003; Suomi and Käyhkö, 2012). The heat island and large momentum transfer cause a deeper boundary layer than over rural areas (Figure 3, Angevine et al., 2003). Furthermore, the atmospheric stability tends to unstable due to Q_F , and the high u_* suppresses the stability towards neutral (due to elevated L). Consequently, the stability is often slightly unstable even in a high-latitude city (**Paper IV**).

These large urban–rural surface exchange differences have motivated EC measurements also in urban environments during the past decade. The majority of research is directed to cities at mid-latitudes (Nemitz et al., 2002; Christen and Vogt, 2004; Grimmond et al., 2004; Piringer and Joffre, 2005; Rotach et al., 2005; Moriwaki and Kanda, 2006; Offerle et al., 2006; Pigeon et al., 2007; Lemonsu et al., 2008; Bergeron and Strachan, 2010; Weber and Kordowski, 2010; Wood et al., 2010), but recently measurements have been conducted also at lower latitudes (Frey et al., 2011) and higher latitudes (Vesala et al., 2008b; **Papers II** and **IV**). Measurements over snow-covered cities (Bergeron and Strachan, 2010) and during stable stratification (Weber and Kordowski, 2010; Wood et al., 2010) are especially scarce.

2.3.2 CO₂ fluxes

In addition to energy exchange, inland waters and urban areas interact with the atmosphere via CO₂ fluxes (F_c). The CO₂ budget of a certain surface consists of its sources ($F_c > 0$) and sinks ($F_c < 0$). The CO₂ exchange of lake surfaces has been studied elsewhere (Anderson et al., 1999; Morison et al., 2000; Eugster et al., 2003; Vesala et al., 2006; Guerin and Abril, 2007; Jonsson et al., 2008; Huotari et al., 2011) and is not addressed here.

The CO₂ balance of a city consists of anthropogenic sources in addition to natural sources and sinks: fossil fuel combustion in housing, transportation and industry is the main anthropogenic source; vegetation can act as a sink through day-time photosynthetic uptake or as a source through night-time respiration (Velasco and Roth, 2010; **Paper III**). In summary, cities are major net sources of CO₂ and are thus hotspots in climate-change mitigations (IPCC, 2007; Kennedy et al., 2009).

CO₂ fluxes can be directly measured using the EC technique, and as a result of the need of direct CO₂ budget measurements, several long-term F_c measurements have

been published within the past two years (Bergeron and Strachan, 2011; Christen et al., 2011; Crawford et al., 2011; Helfter et al., 2011; Hiller et al., 2011; Pawlak et al., 2011; Gioli et al., 2012; Song and Wang, 2012; **Paper III**). In addition to general CO₂ flux dynamics, annual budgets of F_c are readily reported, and annual F_c budgets have been shown to depend exponentially on the fraction of natural land in the EC footprint (Nordbo et al., 2012). The calculation of annual budgets is generally hindered by the data coverage of the EC technique: commonly about 30 % of F_c data are rejected due to quiescent turbulent conditions and more is missing due to flow distortion or instrument malfunction. As a consequence, the choice of a flux data gap-filling method is crucial when evaluating annual F_c budgets, but none of the published work has evaluated the effect of different gap-filling methods.

2.4 Challenges in surface–atmosphere interaction studies using the eddy-covariance technique

The current challenges in the application of the eddy-covariance technique in studying surface–atmosphere exchange can be divided into four categories. (i) Challenges regarding equipment development are mainly related to the accurate and fast response measurements of gases with a lower atmospheric concentration, such as CH₄ (Peltola, 2011) and N₂O (Mammarella et al., 2010; Werle, 2011). More abundant gases, H₂O and CO₂, can be readily measured with infrared gas analysers (Burba et al., 2010), although H₂O concentration fluctuation measurements suffer from sorption. (ii) The calculation methodology behind fluxes and their footprints develops continuously, but a consensus on the optimal methods has not yet been reached, and new calculation problems arise as equipment develops (Aubinet et al., 2012). (iii) The applicability of the EC technique at complex sites or on high towers (>100 m) is questionable due to the inherent assumptions of the technique (introduced in section 2.2). It is also desirable to have a coverage of measurements for all possible ecosystems, not only forests and grasslands that have been at the frontier of EC measurements. (iv) The application of EC measurements at a local to global level to improve weather forecasts (Boussetta et al., 2012) and to determine greenhouse-gas budgets (Schulze et al., 2009; Beer et al., 2010; Nordbo et al., 2012). Wide networks, including open data repositories, for measurements over vegetative surfaces have been established (Valentini et al., 2000; Baldocchi et al., 2001; Mizoguchi et al., 2009) but such networks for measurements over urban

([Grimmond and Christen, 2012](#)) and water surfaces are still emerging.

The main focus of this thesis is on challenges (ii) and (iii). Existing calculation procedures are compared in order to evaluate methodologies at an urban site (**Paper II**), since such a study is still lacking. A new spectral correction method that does not assume spectral similarity is developed (**Paper V**). The long-term energy balance of a small boreal lake is determined, since so far only studies at large lakes or of short-term have been conducted (**Paper I**). Energy balance components at three sites within a northern city are evaluated, which covers a gap in knowledge on the energy exchange in high-latitude cities and within one city (**Papers II** and **Paper IV**). The annual CO₂ budget at a semi-urban site and its inter-annual variability are studied together with the evaluation of gap-filling methods, in order to respond to the need of multi-year analysis of CO₂ budgets and errors in gap-filling methods (**Paper III**).

3 Sites and measurements

3.1 Lake Valkea-Kotinen

The data presented in **Paper I** were collected at Lake Valkea-Kotinen, Southern Finland (61°14'31.20"N, 25°3'48.97"E, 156 m a.s.l., Figure 6a). The lake is relatively small and shallow with an area of 0.041 km² and maximum depth of 6 m. The water is turbid and has a high amount of dissolved organic carbon with a diffuse attenuation coefficient of 3.1 m⁻¹ (Arst et al., 1999) and an average concentration of carbon of 13.5 mg l⁻¹ (Huotari et al., 2009). As a result, the lake acts as a weak net source of CO₂ to the atmosphere (0.077 kg C m⁻² yr⁻¹, Huotari et al., 2011).

The lake was monitored during ice-free periods of 2003–2009, including different spans of micrometeorological measurements. An eddy-covariance station, which provides u_* , H , LE and F_c , was in place throughout the period. Auxiliary meteorological measurements were running until the end of 2007 and net radiation measurements were carried out between 2006 and 2009. The water temperature profile was measured continuously throughout the period and enabled the calculation of ΔQ_S . The meteorological and flux measurements were performed on a raft located near the centre of the lake. Wind direction screening was needed to discard data representing the surrounding forest, but this did not amount to much since the airflow is mainly channelled along the lake. Footprint analysis and first measurements at the lake during the open-water period of 2003 are reported in Vesala et al. (2006), whereas **Paper I** concentrates on the energy balance during open-water periods 2005–2008.

3.2 Helsinki

The work in **Papers II**, **III** and **V** is based on measurements at the SMEAR III site, Helsinki (60°12.17'N, 24°57.67'E, Figure 6b, Järvi et al., 2009a; Järvi et al., 2009b). The station is one of the sites in the SMEAR series (Station for Measuring Ecosystem-Atmosphere Relationships) and consists of four parts: (i) tower-based measurements of temperature, wind speed, net radiation and eddy-covariance fluxes, (ii) roof-top measurements of basic meteorology, (iii) gas and aerosol concentration measurements and (iv) tree physiology measurements. Components (i)-(iii) are at the Kumpula campus

of the University of Helsinki 4 km north-east from the city centre, whereas component (iv) is based at the Viikki campus, about 4 km north-east from Kumpula.

The eddy-covariance measurements were established in 2005 and SMEAR III is the World’s northernmost urban flux measurement station. The EC setup is mounted on a 31-m-high lattice tower and consists of a sonic anemometer and two IRGAs (open- and closed-path). The area surrounding the EC tower can be divided into three wind direction sectors—built, road, vegetation—designated according to dominant land-cover (Vesala et al., 2008a, Figure 7). A heavily trafficked road with 44 000 vehicles per workday passes the tower to the east (Lilleberg and Hellman, 2011). Overall, SMEAR III is characterized by a high fraction of natural area (45%) and will be referred as the semi-urban site in the following.

The measurements used in **Paper IV** were conducted at the SMEAR III station and two new urban stations located in downtown Helsinki: the Fire Station and Hotel Tornii (Figure 6c and d, Figure 7). Both new sites have an EC setup on a mast mounted on a tall building with a measurement height of about 42 m and 60 m, respectively (Figure 2 in **Paper IV**). The 75% footprint contribution lies within 500 m and 800 m from the tower, respectively, for prevailing stability conditions (Figure 7; Figure 1 in **Paper IV**). The surrounding areas are highly built-up with a low fraction of natural area (< 7%) and an average building height exceeding 20 metres. The traffic rate downtown is half of that near SMEAR III (Lilleberg and Hellman, 2011) since the downtown area is characterized by a steadily distributed traffic flow instead of a heavily trafficked road. The Fire Station site was operational July 2010 – January 2011, whereas the Hotel Tornii site was established in September 2010 and measurements are intended to be long-term. More detailed information on instrumentation and surface-cover for all sites are given in Table 1 in **Paper IV**.

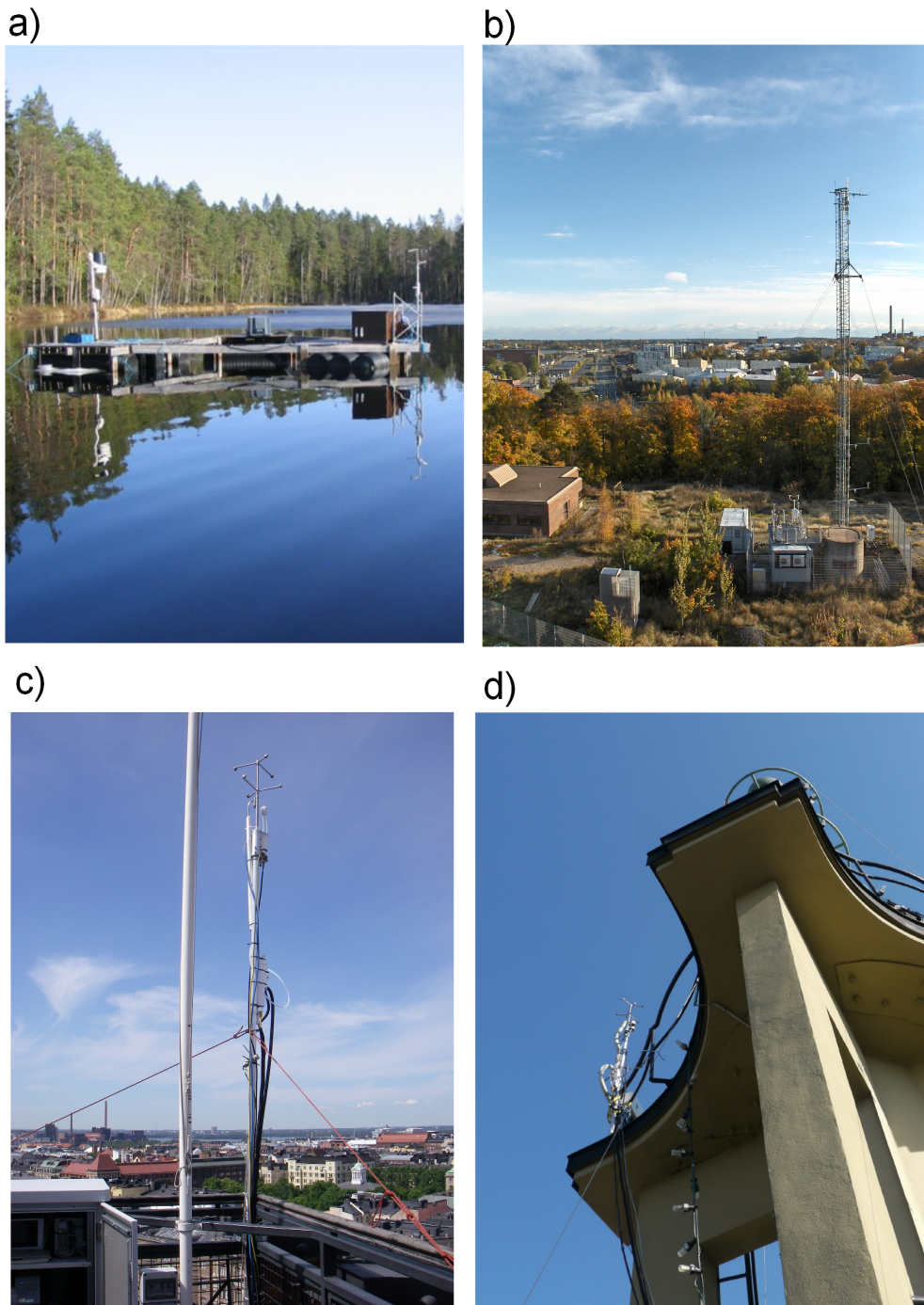
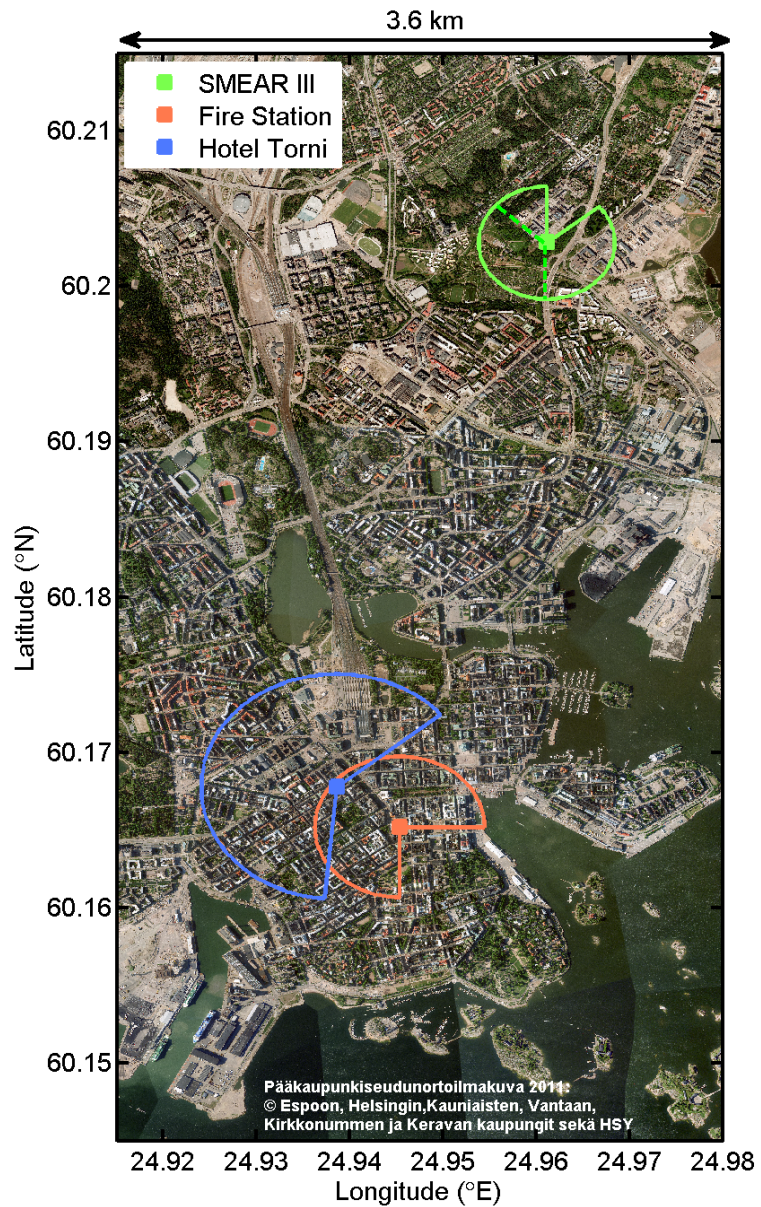


Figure 6: Photographs of measurement sites: a) Lake Valkea-Kotinen (**Paper I**, photo: Timo Vesala), b) SMEAR III (Helsinki, **Papers II, III, IV, V**, photo: Antti-Jussi Kieloaho), c) Fire Station (Helsinki, **Paper IV**), d) Hotel Tornii (Helsinki, **Paper IV**, photo: Janne F. J. Korhonen)

Figure 7: Aerial photo of Helsinki with the three eddy-covariance measurement stations and sectors without flow distortion: SMEAR III (radius 400 m), Fire Station (500 m) and Hotel Tornii (800 m). Circle radii show approximately the 75% flux contribution area for the prevailing stability of the site. Three land-cover sectors are depicted with dashed lines for SMEAR III: built 320–40°, road 40–180° and vegetation 180–320°.



4 Overview of key results

4.1 Energy fluxes at Lake Valkea-Kotinen and three urban sites in Helsinki

At Lake Valkea-Kotinen, R_n and ΔQ_S dominate the energy balance, and diurnal midday maxima exceed 400 W m^{-2} during summer (Figure 8; Figures 8 and 9 in **Paper I**). The cumulative heat storage reaches its maximum in July–August and is highly determined by the timing of final ice-out in the spring (Figure 10 in **Paper I**). H gets its maximum values in early morning, whereas LE peaks in the afternoon (Figure 8), as observed at other lakes (Spence et al., 2003; Liu et al., 2009b). LE dominates over H , but diurnal courses of both stay below 120 W m^{-2} . Substantial nocturnal evaporation, which does not occur at vegetated areas, is also observed to be caused by water heat loss ($\Delta Q_S < 0$). An analysis of the driving forces behind fluxes shows that 62% of the variation in H can be explained by wind speed and surface water–air temperature difference ($U(T_s - T_a)$) and 70% of that of LE can be explained by surface water–air vapour pressure difference ($e_s - e_a$) (Figure 7 in **Paper I**), which has implications for lake energy balance modelling.

The annual mean energy balance closure $\left(\frac{H+LE}{R_n-\Delta Q}\right)$ at Lake Valkea-Kotinen is 72% and 82% for 2007 and 2008, and corresponds to those observed over vegetative surfaces (Wilson et al., 2002). Similar long-term energy balance closure studies are not available from lake sites, but short-term measurements indicate similar closures (Blanken et al., 2000; Liu et al., 2009b). No particular reason for the loss of energy is identified and there is no explained variation in the monthly energy balance closures, with respect to meteorological conditions or ice phenology. Moreover, only the water above the thermocline is shown to contribute to the intra-daily energy balance of the lake (Figure 4 in **Paper I**). The thermocline develops in the lake water a few weeks after ice-out, deepens towards autumn until a turnover takes place around September (Figure 3 in **Paper I**). The decoupling of deeper layers also causes the lake water to heat up in the spring 25% slower than to release the heat in the autumn.

The energy flux partitioning for cities departs largely from that of a lake: at SMEAR III, H dominates over LE , except for summertime vegetation, and ΔQ_S is almost negligible (**Paper II**). The relative magnitudes of H and LE depend largely on wind direction due to a varying surface cover within the footprint (Figure 5 in **Paper II**, Vesala et al., 2008a), and the diurnal peaks of all the energy balance components

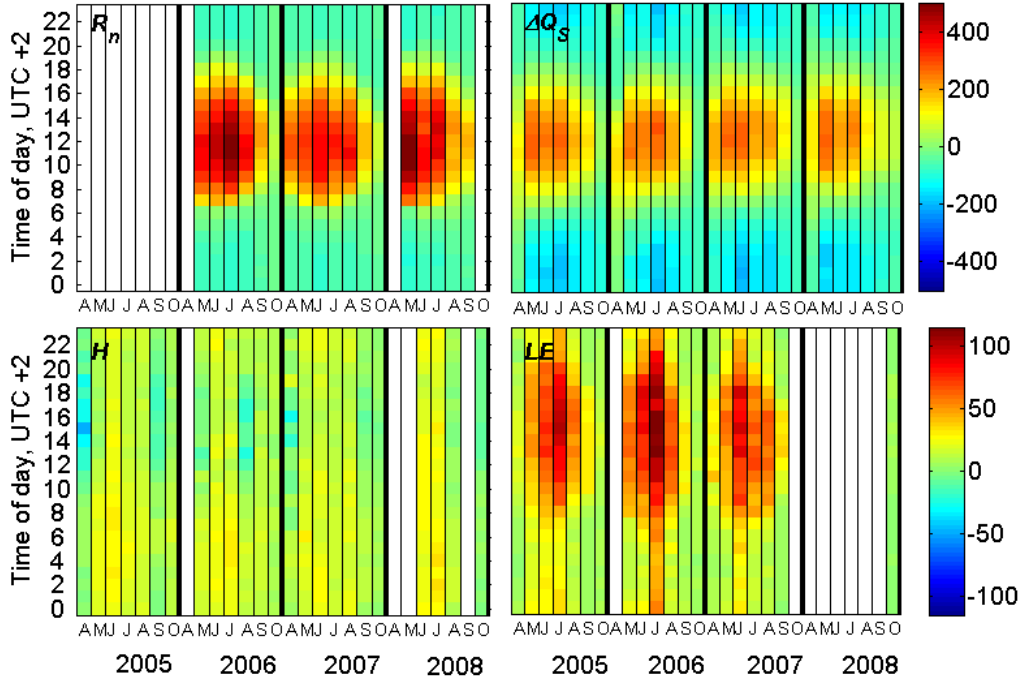


Figure 8: Fingerprints of energy fluxes at Lake Valkea-Kotinen: average diurnal courses per month (Apr–Oct) for the open-water periods of 2005–2008. R_n – net radiation, ΔQ_S – change in water heat storage, H – sensible heat flux, LE – latent heat flux. The flux magnitudes are given in colourbars in W m^{-2} . The figure is produced based on data in **Paper I**.

almost coincide after midday (Figure 9). The monthly mean energy balance residual at SMEAR III varies between -70 W m^{-2} in winter to $+20 \text{ W m}^{-2}$ in summer (Figure 10 in **Paper II**), indicating that heat storage exceeds the anthropogenic heat release ($\Delta Q_S > Q_F$) in the summer. If the annual ΔQ_S is assumed to be zero, the annual mean Q_F around SMEAR III can be estimated to be about 13 W m^{-2} , which is close to values observed in other cities (Table 4 in **Paper II**).

The turbulent fluxes of sensible and latent heat vary across the city. All three sites in Helsinki were running simultaneously October in 2010 – January 2011, and median diurnal cycles show that nocturnal H is about 50 W m^{-2} larger downtown compared with the semi-urban site (Figure 9). The elevated H indicates excess Q_F and leads to a prevailing unstable atmospheric stratification ($\zeta < -0.01$) in central Helsinki compared with the neutral stratification ($-0.01 < \zeta < 0.01$) at SMEAR III (Figure 6 in **Paper IV**). Moreover, LE is smaller downtown due to a lower vegetation fraction.

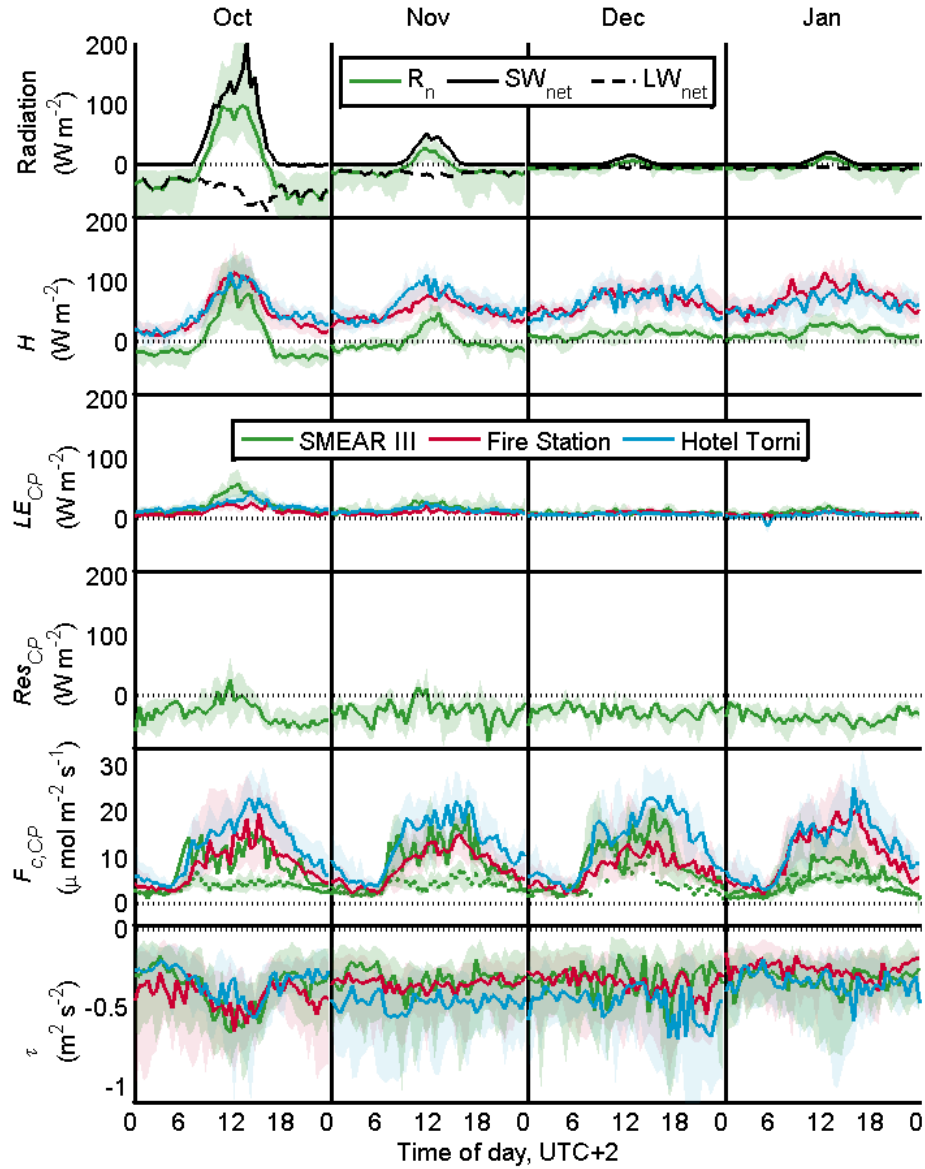


Figure 9: Median monthly diurnal courses of turbulent fluxes for SMEAR III (green), Fire Station (red) and Hotel Tornii (blue) for the period when measurements overlap (October 2010 – January 2011). Columns are months and rows are fluxes: Radiation (R_n net radiation, SW_{net} short-wave net radiation, LW_{net} long-wave net radiation), sensible heat (H), latent heat (LE), energy balance residual (Res), CO_2 flux (F_c), momentum flux (τ). Shading shows 25th and 75th percentiles. Two curves are given for F_c at SMEAR III: flux from the road sector (solid line) and from the vegetation sector (dashed line). R_n and thus Res are only available for SMEAR III. Figure is modified from [Paper IV](#).

4.2 Morphology and turbulence structure in Helsinki

A morphology-based method for determining z_0 and z_d (MacDonald et al., 1998) was improved to take into account the direction from which flow approaches the EC tower (Paper IV). The morphological determination of surface roughness is superior to rule-of-thumb estimates (14% and 44% departure from logarithmic wind profile, respectively), and z_d is estimated to be about 14.6 m and z_0 1.3 m in the centre of Helsinki (Table 1 in Paper IV). Moreover, a fit representing the parameterization of the drag coefficient ($C_D = u_*^2/U^2$) as a function of the ratio of measurement height to mean building height (z/z_H) was revised using data from over 20 sites and some wind tunnel data (Figure 10b).

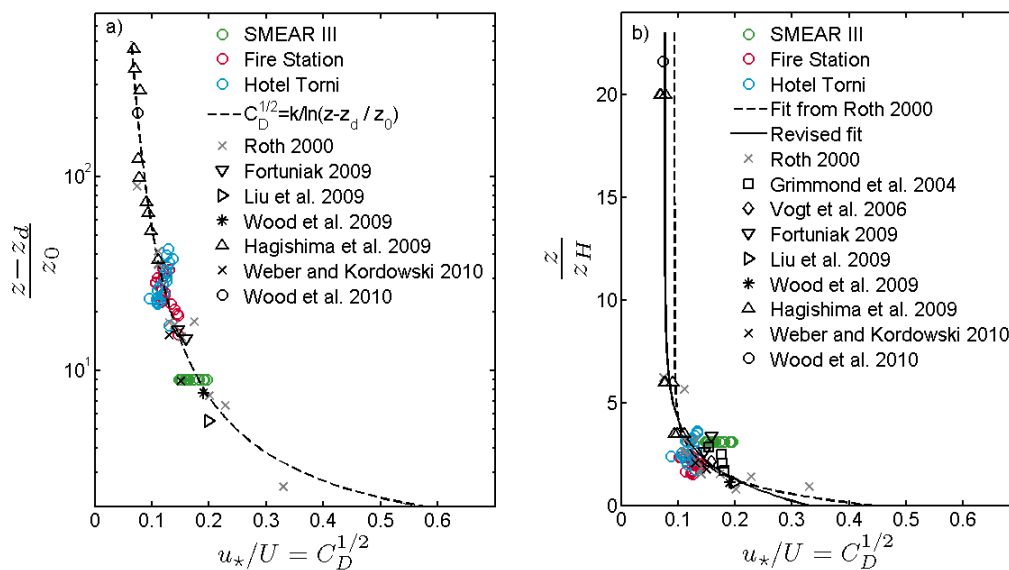


Figure 10: Variation of a) $(z - z_d)/z_0$ and b) z/z_H as a function of the square root of the drag coefficient ($C_D^{1/2}$). In subfigure a), the dashed curve is the logarithmic wind profile in neutral stratification. In subfigure b), the dashed curve follows an empirical form by Roth (2000) and the solid curve is a revised fit including the data with black markers. The data points from the Helsinki sites are averages of 20 bins. Modified from Figure 7 in Paper IV. (Roth, 2000; Grimmond et al., 2004; Vogt et al., 2006; Fortuniak, 2009; Hagishima et al., 2009; Liu et al., 2009a; Wood et al., 2009; Weber and Kordowski, 2010; Wood et al., 2010)

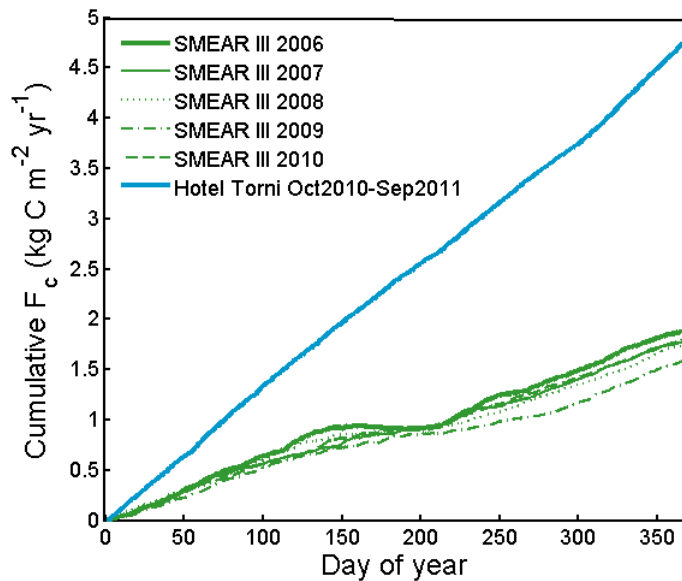
The structure of turbulence over varying urban surfaces was studied in terms of spectra, integral turbulence characteristics and turbulent-transfer efficiencies. All power and co-spectra exhibit clear peaks which indicates that the measurements were conducted above the roughness sublayer introduced earlier in Figure 3. The momentum co-spectra are distorted in that they have a $-3/3$ (instead of $-4/3$) decay in the inertial subrange and a gap around the eddy size of a few hundred metres (Figure 10 in **Paper IV**). The same gap is also observed in scalar transport, but stronger transport at the lowest frequencies is also apparent. Furthermore, the integral turbulence characteristics for horizontal wind speed do not show a dependency on ζ in unstable stratification due to the contribution of large-scale turbulence. The characteristics of CO_2 and H_2O transfer also showed independence of ζ under stable conditions, as also discovered for Beijing (Quan and Hu, 2009). The transport efficiency of heat, H_2O and CO_2 , conversely, is weaker for more heterogeneous sites, and scalar similarity does not apply at our three urban sites probably due to the heterogeneous distribution of sources and sinks (Figure 9 in **Paper IV**). Furthermore, the multi-site measurements also enable the evaluation of horizontal scales that are relevant for turbulent transport. Roughness characteristics (z_0 and z_d) vary already at a ten-metre scale, C_D at a hundred-metre scale, ζ varies markedly within 4 km in Helsinki and normalized power and co-spectra do not vary across Helsinki, provided that appropriate normalization has been performed, i.e. z_d has been calculated correctly.

4.3 CO_2 fluxes in Helsinki

Long-term CO_2 flux measurements at SMEAR III were analysed in **Paper III** and a short-term multisite analysis was performed in **Paper IV**. The CO_2 emissions, based on the eddy-covariance measurements, in central Helsinki peak around 3 p.m. local time, whereas two rush-hour peaks are seen at SMEAR III (Figure 9; Figure 6 in **Paper III**). F_c is greatest at Hotel Tornii (median values exceeding $20 \mu \text{ mol m}^{-2} \text{ s}^{-1}$) followed by Fire Station and the road sector of SMEAR III. The vegetation sector of SMEAR III is a net sink of about $9 \mu \text{ mol m}^{-2} \text{ s}^{-1}$ during summer days, and respiratory emissions from soil are about $5 \mu \text{ mol m}^{-2} \text{ s}^{-1}$, depending on the season (Figure 10 in **Paper III**). The role of vegetation can also be seen as a summer-time plateau in the annual cumulative sum of F_c at the semi-urban SMEAR III site, but the emissions at Hotel Tornii have a continuous level throughout the year (Figure 11).

The mean annual CO₂ budget at SMEAR III is 1.8 kg C m⁻² yr⁻², which classifies the site’s emissions average among the other few studies that are currently available (Figure 11; Table 4 in **Paper III**; Nordbo et al., 2012). The road sector’s net emissions are four times higher than the vegetation sector’s (3.5 and 0.87 kg C m⁻² yr⁻²), and the net emissions are closer to those observed in central Helsinki (4.7 kg C m⁻² yr⁻² at Hotel Tornii, Nordbo et al., 2012). The central Helsinki observations are only half of the inventory-based estimates for the whole of Helsinki (10.8 kg C m⁻² yr⁻²; Heinonen and Junnila, 2011). The discrepancy is caused by the scope difference: the inventory-based estimate includes emissions outside the city (e.g. aviation, electricity production) and excludes the uptake by vegetation. To the contrary, the EC-based budgets show large emissions compared with the uptake of European grasslands (-0.24 ± 0.07 kg C m⁻² yr⁻²; Soussana et al., 2007) or the emissions from Lake Valkea-Kotinen (0.077 kg C m⁻² yr⁻¹; Huotari et al., 2011).

Figure 11: Cumulative sum of the CO₂ flux (F_c , kg C m⁻² yr⁻¹) at SMEAR III for 4 years and Hotel Tornii for one year. SMEAR III data have been gap filled using artificial neural network techniques and Hotel Tornii data with median diurnal cycles. SMEAR III data are adapted from **Paper III** and Hotel Tornii data from Nordbo et al., 2012.



4.4 Systematic errors and random uncertainties in urban flux calculations and gap-filling

Systematic errors in eddy-covariance calculations can originate from a variety of steps (Figure 12). Omitting spectral corrections underestimates LE_{CP} (CP for closed-path IRGA) at SMEAR III by 26%, due to sorption effects of H_2O with tube walls. Conversely, τ , H and LE_{OP} are underestimated by about 5%. Making a planar fit for the flow coordinate rotation underestimates all fluxes by about 5% and the effect is highly stability-dependent. Omitting the surface heating correction of open-path IRGAs causes a small ($< 2\%$) underestimation in LE_{OP} , but the effect is known to be larger for F_c (Järvi et al., 2009b). Insufficient tube-travel time in closed-path measurements causes a 15% lack of data in LE_{CP} , because the lag time is observed to depend exponentially on air relative humidity (Figure 7 in Paper II, Clement, 2004; Ibrom et al., 2007b; Runkle et al., 2012). This is again due to the sorption processes H_2O undergoes in the sampling tube. The lag time also depends on the frequency of variation (Massman and Ibrom, 2008), but this has not yet been verified by measurements. The existence of the sorption processes can also be seen in the cross-correlation functions between w' and water vapour fluctuations (Figure 8 in Paper II). As a result, deficient eddy-covariance calculation procedures may affect the urban energy balance residual by a factor comparable to the magnitude of the residual itself.

Random uncertainties in EC measurements can be calculated in several ways from which none is dominant (Billesbach, 2011). The relative random uncertainty of eddy-covariance measurements (Lenschow et al., 1994) in an urban environment range on average from 9% to 25%, depending on the flux and site in concern (Papers II, III): the relative uncertainty is lowest for H and highest for F_c and it increases the weaker the flux. The detection limit of all fluxes was 10% to 13% of the measured flux, indicating that the fluxes were strong enough to be detected by the EC setups.

The gap-filling of F_c data may cause a bias in the annual budget estimates and thus two gap-filling methods were tested: artificial neural networks (e.g. Moffat et al., 2007) and median diurnal courses (Paper III). The biased error is always below 5% for both tested gap filling methods and for all land-use types at SMEAR III (Table 3 in Paper III). The error is largest for the road sector, and the inclusion of traffic rates does not improve the performance of the artificial neural networks. Interestingly, the median diurnal cycles are shown to give as good results as artificial neural networks in gap-filling of F_c measurements (1% difference).

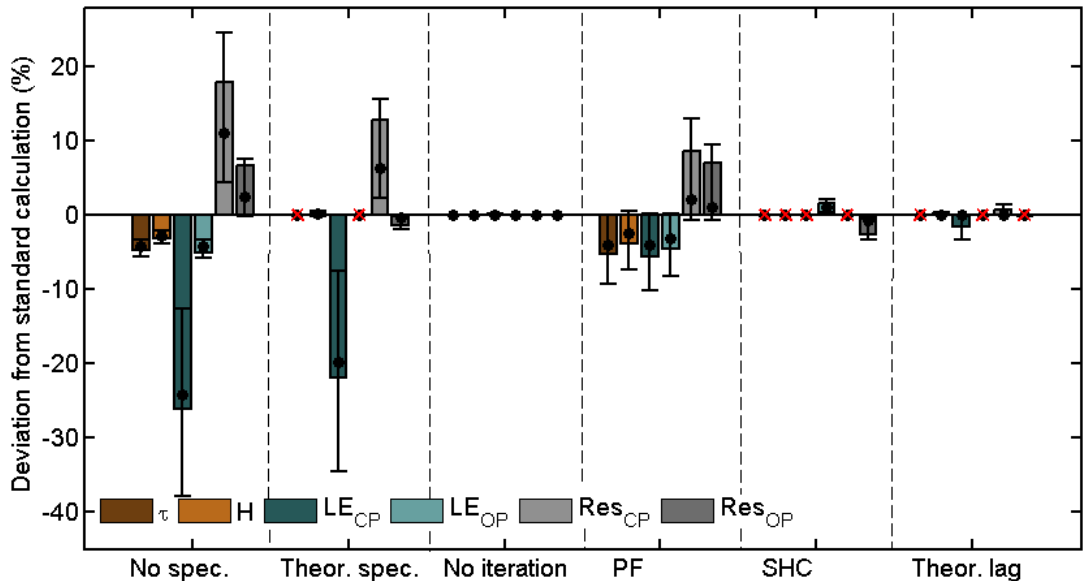


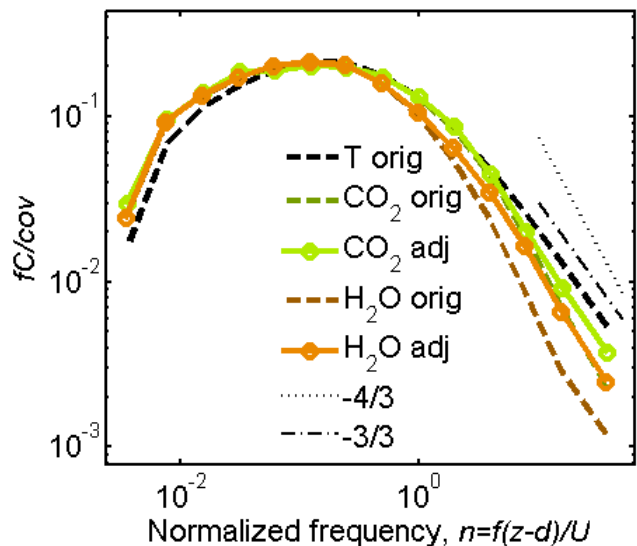
Figure 12: Calculation-procedure effects on turbulent flux magnitudes for momentum (τ), sensible heat (H) and closed- and open-path latent heat (LE_{CP} , LE_{OP}), in addition to effects on energy balance residuals (Res_{CP} , Res_{OP}). Bars give the means, circles (\bullet) denote medians and the errorbars denote 25th and 75th percentiles. Red crosses show fluxes that are not affected by a certain calculation procedure. Deviations are calculated as $(\text{flux} - \text{flux}_{\text{Standard}}) \text{flux}_{\text{Standard}}^{-1}$ except for Res where an absolute value of the denominator is used, since corrections can easily change the sign of Res . The calculation versions on the x-axis are: no spectral corrections, theoretical spectral corrections, no iteration, planar fitting, surface heating correction and theoretical lag window. See **Paper II** and its Figure 6 for further details.

4.5 A new wavelet-based spectral-correction method

As shown above (Figure 12), the high-frequency spectral corrections are the most crucial calculation step, especially for H_2O flux measurements with a closed-path IRGA. Thus, a new method to account for high-frequency losses in eddy-covariance measurements was developed (**Paper V**). It relies on a different set of assumptions than the previously introduced theoretical and experimental approaches, and it is fully automatic.

In the method, wavelet analysis is used for re-creating small-scale variation in raw data (e.g. 10 Hz) scalar time series. The recreation is done by using information of the peak location and inertial subrange slope of instantaneous power spectra, adjusting the time series in wavelet space and performing an inverse transform to retrieve a new time series. The adjusted time series is proven to have turbulence properties superior to the original time series (Figure 7 in **Paper V**), and attenuation at high frequencies is not apparent any more in the scalar co-spectra of the adjusted time series (Figure 13; Figure 5 in **Paper V**). The method is also able to capture the relative humidity dependency of tube attenuation of water vapour, and the method is able to link the system response time and tube travel lag time in water vapour flux measurements (Figure 5 and 6 in **Paper V**). The proposed wavelet-based spectral correction method yields corrections that are larger than those from the theoretical approach and smaller than those from the experimental method (Figure 4 in **Paper V**). The larger experimental-based corrections might be due to (i) scalar dissimilarity, which causes an overestimation of the correction and/or (ii) the wavelet-based corrections inability to account for variation loss that starts already around the peak of the scalar power spectrum.

Figure 13: Average normalized cospectra for temperature (T), carbon dioxide (CO_2) and water vapour (H_2O) as a function of normalized frequency. Cospectra are given for original data and wavelet-correction adjusted data, and the theoretical inertial subrange slope of $-4/3$ (traditionally expected for cospectra) and $-3/3$ (heterogeneous situation) are also displayed. Data have been bin-averaged according to normalized frequency. Adapted from Figure 3a in **Paper V**.



5 Review of papers and contributions

This work consists of five publications (Figure 14) that investigate the surface–atmosphere interactions at non-standard sites (a lake, **Paper I**; three urban sites, **Paper II, IV, III**) in addition to the strengths and weaknesses of the eddy-covariance technique (**Paper II, V**).

Paper I is the first long-term energy-balance study based on direct eddy-covariance measurements over a small lake. The work addresses the energy imbalance problem and evaluates the importance of energy-balance components and the importance of the water layer below the thermocline.

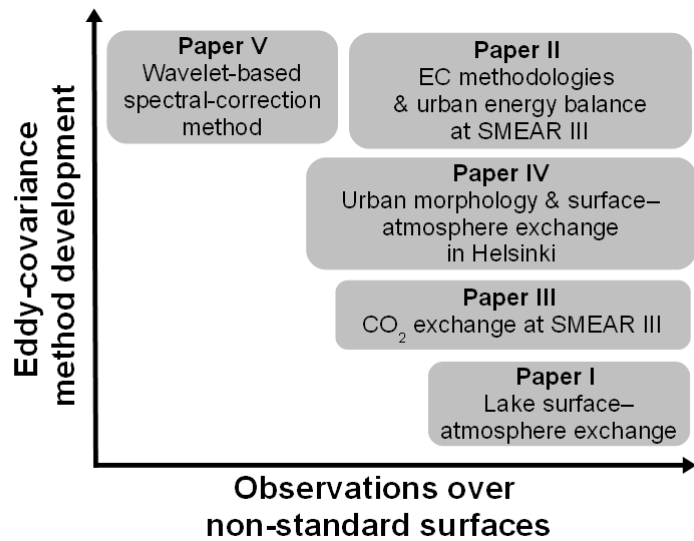
Paper II provides the first comprehensive evaluation of the effect of systematic errors and random uncertainties in EC measurements at an urban site. The effect of systematic errors on energy balance closure and implications on anthropogenic heat release are addressed. Special attention is given to the problems that sorption of water vapour causes to EC flux estimates of latent heat.

Paper III quantifies the annual CO₂ budget for SMEAR III based on five consecutive years of direct eddy-covariance measurements. This study is the first to make a comprehensive analysis of inter-annual variability in the CO₂ budget in addition to the random errors and the effect of gap-filling methods on the final budget value in an urban environment.

Paper IV introduces two new measurement sites in the centre of Helsinki and provides a multi-site analysis of turbulence structure within one city. The measurement setups are thoroughly reported and surface roughness parameters around the measurement sites are determined morphologically. Horizontal scales at which turbulence-relevant quantities vary are analysed, and implications to grid-sizes in urban modelling are made. Dissimilarity in scalar transport and the shapes of urban spectra are studied.

Paper V demonstrates a new wavelet-based method for correcting for the loss of flux transported by the smallest eddies. The method is tested by comparing with other methods and by using structure parameters and turbulence spectra. The method’s ability to automatically adapt to contrasting sites was also examined.

Figure 14: Summary of the contribution of papers in this thesis to: EC technique development vs. measurements over non-standard surfaces. The extent and location of the grey boxes on both axes indicates how the paper concerned contributes to the topics on the axis.



I am solely responsible for the summary part of this thesis and for the final raw data post processing and quality assurance of all eddy-covariance data in the publications. In **Papers I, II, IV and V**, I am responsible for most of the writing and all data analysis (except the surface heating correction in **Paper II** and footprint calculations in **IV**). In all papers, the co-authors contributed to the interpretation and writing of the results. My contribution to **Paper III** is the least out of the five papers in this thesis: I only made the final flux, random error and detection limit calculations, and participated in the interpretation and writing of the results. The measurements in **Paper I** were conducted by the staff of the Department of Environmental Sciences and those in **Papers II-V** by the staff of the Department of Physics at the University of Helsinki. I participated in the installation of measurements related only to **Paper IV**, since all other sites were already implemented prior to the start of this thesis work.

6 Conclusions and outlook

The eddy-covariance technique has been accepted as the conventional way to study surface–atmosphere interactions. Yet, a final consensus on its methodologies, such as coordinate rotations, density corrections and spectral corrections, has not been reached. Furthermore, the technique has not been applied over all surface types to a satisfactory extent in order to understand how different surfaces affect the state of the atmosphere. This thesis has addressed these two issues and the main conclusions are as follows.

The surface–atmosphere exchange in an urban environment is affected by anthropogenic influence: a 1.8 kg C m^{-2} annual emission of CO_2 and a 13 W m^{-2} release of heat are observed at a semi-urban SMEAR III site in Helsinki. The inter-annual variability in the CO_2 balance is small ($< 6\%$ over 5 years), but the difference between land-cover sectors is large (75%). The CO_2 emissions in central Helsinki are comparable to those from a heavily trafficked road in a sector of the semi-urban site. The effect of anthropogenic heat release and changes in surface-cover are consistent with a large difference in atmospheric stability between central Helsinki and the semi-urban site. In the surface–atmosphere exchange in a lake environment, the lake water itself is a determining factor in surface–atmosphere interactions: the water below the thermocline is quantitatively shown to have small importance in intra-daily energy balance, and the change in water heat storage is a major component especially since it is able to fuel substantial nocturnal evaporation.

The energy imbalance problem in EC measurements is not primarily terrain-dependent, since about the same lack of energy balance closure is observed at a small boreal lake compared with vegetated surfaces. The imbalance at any site can also be caused by unintentional errors in EC calculation procedures. At a semi-urban site in Helsinki, these errors can alter the energy balance residual by an amount similar to the residual itself, and the errors propagate further to the evaluation of the anthropogenic heat release (above). Though the exact magnitudes of the effect of specific calculation procedures are setup-specific, the results (Figure 12) can be taken as an example and warning for EC analysts, also in non-urban surroundings.

Water vapour flux measurements with a closed-path analyser are prone to errors due to sorption effects when flowing down a sampling tube: the signal’s arrival is delayed and it is attenuated. If the delay of the signal is not appropriately accounted for, 15% of data may be lost due to data quality screening. A new wavelet-based spectral-correction method is developed to automatically correct for this signal attenuation of

H₂O and other constituents. The method does not assume similarity in scalar transport, which is an advantage since strong scalar dissimilarity was observed at the three sites in Helsinki. Spectral corrections have been the most laborious step in EC flux calculations and are hindering the near-real-time usage of the measurements. The new method provides an initial solution for this automation problem. The work shows the usefulness of wavelet-analysis in conjunction with EC measurements and may encourage development of other wavelet-based applications in EC raw data post-processing.

The results of the lake and urban energy balance studies are applicable at least for small lakes in the boreal zone and high-latitude cities, respectively. The lake data are currently used for improving lake modelling components in weather forecast models and to evaluate the necessity of surface tiling in operational numerical weather forecasting. The urban data will be used in improving the modelling of urban energy fluxes, and the role of anthropogenic heat release will be determined for Helsinki in more detail. The source codes for the wavelet-based spectral correction method are freely available on the internet for download, and lab measurements have been carried out for future research on the sorption effects of H₂O with tube walls.

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