The Pennsylvania State University
The Graduate School
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REGIONAL ESTIMATES OF NET ECOSYSTEM-ATMOSPHERE EXCHANGE
OF CARBON DIOXIDE OVER A HETEROGENEOUS ECOSYSTEM

A Thesis in
Meteorology
by
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Submitted in Partial Fulfillment
of the Requirements
for the Degree of

Doctor of Philosophy

August 2005
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The net ecosystem-atmosphere exchange of CO$_2$ (NEE) is estimated over a mixed forest ecosystem in the 40×40km$^2$ region centered at the WLEF tall tower in northern Wisconsin. Flux aggregation and the atmospheric boundary layer (ABL) budget methods are used. CO$_2$ fluxes measured at three levels of the WLEF tower are decomposed in the growing season to infer the CO$_2$ fluxes for six stand types based on a stand level classification scheme in the tower footprint area. The flux footprint models, vegetation map and eddy-covariance data are combined to estimate the optimal solutions for the parameters in two ecosystem models for each stand type. The results show differences in the parameters and fluxes among the different vegetation classes that are consistent with general expectations for the respective stand types. The inferred fluxes are compared to observations from two stand-level eddy-covariance flux towers, implying that the six stand classification scheme is insufficient to capture all variability in stand characteristics relevant to CO$_2$ exchange. Significant uncertainty in aggregated fluxes exists due to the classification schemes used. In the calculation, a new approximate analytical model is introduced to predict the footprint for flux measurements of passive scalars in the lower part of the mixed layer. An analytical footprint model under the idealized convective boundary layer is adjusted by comparing its solutions to those from a Lagrangian stochastic model with more realistic meteorological conditions.

The ABL budget method is used to infer regional NEE by using the vertical profile of CO$_2$ mixing ratios measured at the WLEF tall tower. The estimated regional NEE is close to that measured at the highest level of the WLEF tower in the day. In the
nocturnal boundary layer, the cases when the atmosphere is very stable or nearly-neutral are screened out due to possible systematic errors. Regional NEE is generally bounded by the current tower measurements. To estimate the effects of horizontal advection on daytime NEE estimates, a water vapor constraint is introduced, where the range of the effects of horizontal advection of CO₂ is estimated by the measurements of the spatial distribution of water vapor. The horizontal advective flux of CO₂, a function of wind direction, is estimated to be smaller than 10-20% of the surface flux in our budget calculation on temporal and spatial scales of one hour and 40km.

Regional NEE estimates from the two methods are compared with each other and with those reported in the literature. In the dormant season, differences among all estimates are relatively small so that NEE measured at the WLEF tower might be an acceptable approximation of NEE in the region (10³ km²) and even in a larger region (10⁴ to 10⁵ km²) on annual or longer time scales. This, however, is not true in the growing season. The cumulative regional NEE values are -103±50 and -175±60gC m⁻² in 2000 and 2003, respectively, suggesting that the region is a net sink of CO₂. Those values are significantly different from those measured at individual towers. It is inappropriate simply to extend fluxes measured at individual towers to a larger region.

**Key words:** footprint modeling, atmospheric boundary layer budget, regional carbon flux estimates, aggregation, flux footprint decomposition
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ACKNOWLEDGEMENTS

I would like to express my deep gratitude to my advisor, Dr. Kenneth J. Davis. The completion of this dissertation would never have been possible without his unfailing patient guidance and continuous encouragement. It is his insightful advice that brings about improvements in the content and structure of the dissertation. Also, it is his academic strength and rigor that guided me not only through every stage of dissertation revision but also through the process of fulfilling my academic goals at Department of Meteorology in The Pennsylvania State University.

My heartfelt thanks also go to Dr. John C. Wyngaard, Dr. Toby N. Carlson, and Dr. Andrew M. Carleton for their constructive suggestions and enlightenment in other fields of atmospheric sciences that are of great help to the academic strength of my dissertation. I benefited a lot from their scholarly expertise and rigorous approach.

I would like to thank all my group members for their help and discussions. To Martha P. Bulter and Daniel M. Ricciuto, I owe a great deal. Without their assistance with the revision of my dissertation, it could not have been where it is today. I also thank Dr. Chuixiang Yi and Bruce D. Cook for their encouragement and discussions. I thank all people for their assistance in acquisition of the data.

Most deeply, I want to thank my parents and my wife for their support and patience. In their love I can always find strength.
Chapter 1

Introduction

1.1 General background of the research

Observations indicate that the mixing ratio of carbon dioxide (CO$_2$) in the atmosphere has been increasing from 280 ppm to 367 ppm over the past century (IPCC, 2001), affecting global and regional climate (Cox et al., 2000; IPCC, 2001; Schlesinger, 1983; Schlesinger, 1997; Tans and Bakwin, 1995). Such a rate of increase, however, is smaller than the rate of anthropogenic emissions, mainly fossil fuel burning and deforestation (IPCC, 2001). The analyses of the global carbon budget suggest that terrestrial ecosystems may play a critical role in buffering the increase (Ciais et al., 1995; Fan et al., 1998; Keeling et al., 1996; Keeling and Shertz, 1992; Tans et al., 1990). The terrestrial carbon flux, however, is the most uncertain among the terms in the global CO$_2$ budget (IPCC, 2001). Causes of spatial and temporal variability of the terrestrial carbon budget are not yet well understood (Conway et al., 1994; Houghton, 2003; Keeling et al., 1996; Kicklighter et al., 1999).

Quantifying terrestrial carbon fluxes and understanding the underlying mechanisms over a variety of spatial and temporal scales are essential to make future projections with more confidence about the impacts of human activities and the resulting
climate change, which is required not only by the scientific community (Huntingford et al., 2000; Tans et al., 1996) but also by the policy community (den Elzen et al., 2005; Tans and Wallace, 1999; UNFCCC, 1997). Limited knowledge about the feedback mechanisms between terrestrial ecosystems and climate change due to the lack of measurements has led to much uncertainty in projections of CO₂ concentration and climate (Cao and Woodward, 1998; Cramer et al., 2001; Cramer et al., 1999; Huntingford et al., 2000; IPCC, 2001; Stephens et al., 1998). Politically, developing emission trading credits and managing the emission at the nation level, e.g., based on Kyoto Protocol (UNFCCC, 1997), also requires an accurate evaluation of the current and future status of carbon fluxes on regional scales, which is challenging the scientific community.

1.2 Current status of carbon flux measurements

Figure 1.1 summarizes the current status of the measurements of net ecosystem-atmosphere exchange of CO₂ (NEE) as a function of spatial and temporal scales. NEE can be measured on a local scale (e.g., 1km²) for a long time period (e.g., years) by the eddy-covariance method that has been widely used in recent decades. Such measurements have been made on over 150 towers, most of which are at forest sites, across the world (Baldocchi et al., 2001). In addition, observed CO₂ mixing ratios and their spatial gradients in the atmosphere have been used to constrain NEE on the global scale by means of inverse modeling (Bousquet et al., 1999a; Bousquet et al., 1999b; Enting et al., 1995; Gurney et al., 2002; Law, 1999; Rayner et al., 1996). NEE, however, is not easily estimated or measured over the regions between local and global scales. Methods for
measuring regional NEE have not been well developed. Although NEE on the scales of the order of 50 to 100 km can be directly observed by aircraft measurement techniques, this method is expensive and is restricted for short time scales such as hours and days (Desjardins et al., 1995; Oechel et al., 1999; Smith et al., 2003; Williams et al., 2001).

In practice, regional-scale NEE can be inferred by bottom-up and top-down methods. The bottom-up method is to extend the measurements or model predictions from smaller spatial scales to larger spatial scales, or from shorter temporal scales to longer temporal scales, or both. A key step in the bottom-up scaling is to properly classify the whole ecosystem into various types according to certain criteria. In doing so, the variability of fluxes (within a type) can be reduced under given environmental conditions, and hence the measurements or model predictions for the fluxes can be more representative of that ecosystem type (Bouwman, 1999). If we can understand the underlying mechanism of the processes of CO₂ exchange and develop process-based models for a specific ecosystem type according to the measurements, it is possible to infer fluxes at new environmental regimes based on the models, which is an advantage of the method. An improper classification and the resulting inadequate representation of the measurements, however, can result in serious errors in the predictions on regional scales because the bottom-up method is open-ended (Jarvis, 1993). For example, the traditional ecosystem classification scheme is based on land-cover types. With this classification scheme, measurements can be made for each ecosystem type, and then can be scaled up to a larger area, e.g., using the aggregation method. Studies, however, suggest that other factors such as plant age and density (Litton et al., 2004; Litvak et al., 2003; Peltoniemi et al., 2004; Wirth et al., 2002) are also important in controlling NEE. Therefore, simple
grouping based on the traditional scheme might not work in some places. Ignoring processes that local measurements cannot observe as well as the uncertainty of the ecosystem distribution map are additional sources of errors, restricting the application of the bottom-up method.

The top-down method uses the mixing ratio of CO\textsubscript{2} in the atmosphere to infer fluxes (Brouwman et al., 1999; Jarvis, 1993), and can test the predictions from the open-ended bottom-up method. Applying this method does not require both the information of the distribution of ecosystems and flux measurements for individual types, which is one of the advantages of this method. The top-down method, however, reflects less mechanism than the bottom-up method. Therefore, the results from the top-down method are more difficult to extend to different situations. Another weakness of the top-down method is that the uncertainty of the results is usually large or cannot be quantified with sufficient confidence. For example, the representation of the mixing ratio measurements or model results is uncertain and some quantities cannot be measured easily. As a result, a combination of the two methods is advantageous.

1.3 Motivations and methods used in the dissertation

1.3.1 Motivations

The study site is located in the Chequamegon National Forest in northern Wisconsin, centered at a tall communication tower (WLEF) (45.9455878°N,
90.272304°W). CO₂ fluxes have been measured using the eddy-covariance method at the WLEF tower, a NOAA-CMDL CO₂ sampling site, since 1995 (Davis et al., 2003). Within an area of 40×40km² centered at the WLEF-TV tower, similar measurements have also been made at two short towers, Willow Creek (WC) and Lost Creek (LC) towers, since 1999 and 2000, respectively. An initial hypothesis was that NEE measured at the WLEF tower could be representative of the region to some extent because the vegetation types in the footprint of the flux measurements are mainly forested uplands and wetlands (Burrows et al., 2002; Mackay et al., 2002). This hypothesis, however, is easily rejected by examining the measurements for a wetland ecosystem and an upland forest ecosystem made at the WC and LC towers, respectively (see Chapter 3). Therefore, it is necessary to figure out why the hypothesis is rejected and its implications for the up-scaling in this region. The other motivation is to estimate NEE with certain confidence in the region, where ecosystems respond differently to the same environmental conditions.

1.3.2 Methods

Methods for regional NEE estimates reported in the literature include airborne eddy-covariance measurements, mass budget method, remote sensing, forward and inverse modeling, isotopic signature method, and others. Recent reviews of the approaches for scaling gas fluxes in ecosystems are given by Bouwman (1999) and Ehleringer and Field (1993). In this dissertation, efforts are made to estimate regional NEE using two independent methods, the flux aggregation and mass budget approaches.
Flux aggregation method is a straightforward way. It is, however, not easy to carry out in practice. At least two questions arise when we conduct the aggregation. First, which level of ecosystem classification is acceptable in terms of both the accuracy of the classification data and the representativeness of the measurements for each ecosystem type? Second, how can the aggregation be conducted for a more detailed land-surface classification level if the direct measurements are available only for a limited number of ecosystem types? In an effort to partially address these questions, we compare the results from aggregation experiments using different levels of classification in a case study.

An independent approach is the atmospheric boundary layer (ABL) budget method, which is based on mass conservation. This method is traditionally only applied to inferring NEE on short time scales (Denmead et al., 1999; Kuck et al., 2000; Pattey et al., 2002) because long-term measurements of the vertical CO\textsubscript{2} mixing ratio profile are usually unavailable. We take advantage of long-term measurements of the CO\textsubscript{2} mixing ratio profile on the 447-m WLEF tall tower so that regional NEE can be estimated on an hourly basis for a time period of years. The results from this method can give another regional NEE estimate. Detailed reviews of both methods are given in the respective chapters.

1.4 Outline of the dissertation

Chapter 2 presents the derivation of an analytical footprint function for flux measurements in an ideal convective boundary layer (CBL) from the Lagrangian perspective. Then the analytical solutions are adjusted to those from a stochastic model
with more realistic meteorological conditions. The adjusted model is used in Chapter 3 for flux decomposition.

Chapter 3 describes the approach to decompose the fluxes measured at the WLEF tower to estimate the fluxes for more ecosystem types in the region. Fluxes for six ecosystem types are inferred in the WLEF footprint area during the growing season in 2000 and 2003 in a case study. Then the inferred fluxes in addition to those measured at two short towers are aggregated in the region. Results suggest that the aggregation based on land-cover types may lead to error in NEE estimates in the region. Limitations and future work are discussed.

Chapter 4 estimates regional NEE values in the day and at night using the ABL budget method. We applied a simple model to predict the CO₂ jump at the top of CBL based on the measurements within the mixed layer and in the lower troposphere. At night, we define criteria to screen the cases when possible systematic errors could occur under very stable or neutral conditions. The results are compared to those measured at towers.

Chapter 5 proposes an approach to estimate the effects of horizontal advection of CO₂ on the daytime NEE evaluations in terms of the measurements of the water vapor mixing ratio and flux. Limitations and further attempts are discussed.

Chapter 6 compares the regional NEE estimates from the two independent methods in Chapters 3 and 4 with each other, and with the results reported in the literature.

Finally, Chapter 7 summarizes the conclusions of this study and discusses future work. Some derivations are given in Appendixes.
Figure 1.1: NEE measurements on temporal and spatial scales. Measurements are usually lacking on the scales of the shaded area.
Chapter 2

An Approximate Footprint Model in the Convective Boundary Layer and Its Application to Flux Measurements at a Tall Tower

2.1 Introduction

Footprint is defined as the contribution of each unit element of the upwind surface area to a measured vertical turbulent flux (Schuepp et al., 1990) using tower-based or aircraft-based eddy covariance. The interpretation of eddy-covariance flux measurements over a heterogeneous surface depends largely on the footprint over which fluxes are sampled. The location and size of this footprint for the measurements at a given height, however, usually vary with meteorological conditions. Therefore, it is crucial for interpreting measured fluxes to estimate the footprint as a function of meteorological conditions.

There are a number of studies of flux footprints for measurements within the surface layer (Horst and Weil, 1992; Schmid, 2002; Schuepp et al., 1990). Models for the footprints have been successfully used to interpret long-term flux measurements within the surface layer (Amiro, 1998; Stoughton et al., 2000). Fluxes have also been measured above the surface layer using aircraft (e.g., Davis et al., 1997; Mahrt, 1998; Oncley et al., 1997) and tall-tower platforms (Davis et al., 2003) in order to cover large horizontal
distances and observe vertical structure within the atmospheric boundary layer (ABL). Unlike within the surface layer, flux footprints for measurements above the surface layer have not been extensively studied. There are numerical investigations in the convective boundary layer (CBL) including a stochastic model (Weil and Horst, 1992), large eddy simulation (Leclerc et al., 1997), and a second-order closure model (Wang and Davis, 2002). None of these methods, however, can be used in practice for long-term calculations. The objective of this study is to construct an explicit analytical model to predict the footprint above the surface layer for interpreting long-term flux measurements in the CBL. In an effort to search for explicit predictions for flux footprints, Kljun et al. (2004) presented a scaling approach that can also simulate footprints for the measurements above the surface layer to the middle of the CBL. They constructed a dimensionless footprint function using the Buckingham $\Pi$ Theorem, in which the parameters are determined by fitting the function to the results simulated from a Lagrangian stochastic model. In contrast to their dimensional analysis method, the present study starts with deriving an analytical solution for the footprint in an idealized CBL. Then the ideal solution is adjusted to major features of the footprint fitted from a Lagrangian stochastic model like others done in the literature (Hsieh et al., 2000; Kljun et al., 2004) due to very limited field data.

Section 2.2 describes the main features of the footprints in the CBL simulated from a Lagrangian stochastic model. The analytical footprint function in an ideal CBL and its adjustment are presented in section 2.3. As an application, we compare the footprints of the fluxes measured within and above the surface layer at a 447-m tall tower.
(WLEF) in northern Wisconsin in section 2.4. Finally, summary and conclusions are given in section 2.5.

2.2 Footprints evaluated from a Lagrangian stochastic model

2.2.1 Model description

The scalar flux footprint, \( f(x, y, z_m) \), is equal to the vertical flux downwind of a unit surface point source (Horst and Weil, 1992; Horst and Weil, 1994), i.e.,

\[
f(x, y, z_m) = \frac{F_m(x, y, z_m)}{Q},
\]

where \( x \) and \( y \) are the horizontal distances of the measurement point from a surface point source (sink) with an emission rate \( Q \) at the origin; \( z_m \) is the measurement height; \( F_m(x, y, z_m) \) is the vertical flux measured at position \( (x, y, z_m) \) in the Eulerian field. In Lagrangian models, \( F_m \) can be evaluated by recording the trajectories of the particles released from the source into the atmosphere. In the fluid with zero mean vertical velocity, \( F_m \) can be written as (van Dop et al., 1985),

\[
F_m(x, y, z_m) = \langle w(x, y, z_m) \rangle C(x, y, z_m),
\]

where \( \langle w(x, y, z_m) \rangle \) is an average of the vertical velocities of the released particles passing the infinitesimal volume \( dxdydz \) centered at \( (x, y, z_m) \); \( C(x, y, z_m) \) is the mean mixing ratio at position \( (x, y, z_m) \), which can be evaluated by counting the number of trajectories passing the volume.
The crosswind-integrated footprint (CWIF), usually examined in the literature, is defined by

\[ f^y(x, z_m) = \int_{-\infty}^{\infty} f(x, y, z_m) dy , \quad (2.3) \]

is then

\[ f^y(x, z_m) = \frac{F^y_m(x, z_m)}{Q} , \quad (2.4) \]

where \( F^y_m(x, z_m) \) is the crosswind-integrated mean vertical flux that can be readily obtained by rewriting Eq. (2.2) in the two-dimensional case (see appendix A.1),

\[ F^y_m(x, z_m) = \langle w(x, z_m) \rangle > C^y(x, z_m) , \quad (2.5) \]

where \( C^y(x, z_m) \) is the crosswind integrated mixing ratio.

In this study, a Lagrangian particle model developed by Luhar and Britter (1989) and modified by Luhar et al. (1996) is selected to evaluate the crosswind-integrated vertical flux of the passive scalar released from a surface point source in the CBL, and hence the footprint. The model has been used successfully to simulate the dispersion in the CBL. The results from the model agree well with those from both water tank experiments (Willis and Deardorff, 1976; Willis and Deardorff, 1978; Willis and Deardorff, 1981) and field observations (Briggs, 1993a; Briggs, 1993b). The model is briefly described as follows.

A large number of particles are released at the source and their trajectories are recorded. The vertical velocity, \( w_p \) and position, \( (x_p, z_p) \), at time \( t \) of a particle are determined by the following equations (Luhar and Britter, 1989; Thomson, 1987),
where $C_0$ is a universal constant, $e$ is the turbulent kinetic energy dissipation rate, and $d\zeta$ is a Gaussian random forcing with zero mean and variance $dt$; $U(z)$ is the mean horizontal wind speed in the mean wind direction; the deterministic acceleration term $a(w_p, z_p)$ is a function of turbulence statistics and is derived from the Fokker-Planck equation incorporating the well-mixed condition (Thomson, 1987). In deriving $a(w_p, z_p)$, the study assumes a skewed probability distribution function (PDF) of $w_p$ parameterized by the sum of two Gaussian PDFs each of which has a nonzero mean velocity. The effects of horizontal velocity fluctuations are ignored, and thus the streamwise displacement is assumed to depend only on the mean wind speed profile. Details can be found in the original reference (Luhar and Britter, 1989).

We slightly modify the Luhar and Britter’s model by considering the inhomogeneity of the mean horizontal wind profile $U(z_p)$ in the surface layer instead of the uniform wind speed in all levels. $U(z_p)$ is estimated using the Monin-Obukhov surface layer theory below $0.1\, h$, where $h$ is the depth of the CBL, and is assumed to be uniform with height above $0.1h$. Because the structure of the CBL has been better documented under strong convective conditions than under weak unstable conditions (Weil, 1988),
only the former conditions, i.e., \(-h/L > 10\), where \(L\) is the Monin-Obukhov length, are investigated. The vertical profiles of \(\overline{w^2}\) and \(\overline{w^3}\) (Stull, 1988) used are,

\[
\frac{\overline{w^2}}{w^*_{\alpha}} = 1.8 \left( \frac{z}{h} \right)^{3/2} \left( 1 - 0.8 \frac{z}{h} \right), \tag{2.9}
\]

\[
\frac{\overline{w^3}}{w^*_{\alpha}} = 0.8 \frac{z}{h} \left( 1 - 1.1 \frac{z}{h} \right), \tag{2.10}
\]

where \(w_*\) is the convective velocity scale. In the surface layer, we use

\[
\left( \overline{w^2} \right)^{1/2} = 1.9u_* \left( -\frac{z}{L} \right)^{1/3} \quad \text{and} \quad \overline{w^3} = 2.5u_*^3 \left( -\frac{z}{L} \right) \quad \text{(Hunt et al., 1988)},
\]

where \(u_*\) is the friction velocity which is related to \(w_*\) by \(w_*/u_* = (-h/L)^{1/3}\), and \(\kappa\) is the von Karman constant. The expressions for \(\overline{w^2}\) and \(\overline{w^3}\) result in a skewness \(S=0.36\) in the surface layer consistent with observations (Wyngaard, 1988). In the surface layer, the \(\varepsilon\) is assumed to be

\[
\varepsilon = \left( u_*/\kappa z \right) \left[ 1 + 0.5 \left( -\frac{z}{L} \right)^{2/3} \right]^{3/2} \quad \text{(Wyngaard and Cote, 1971)}.
\]

Entrainment at the top of the CBL is not considered.

For a surface point source, the mean cross-wind integrated mixing ratio \(C^x(x, z_m)\) is given by,

\[
C^x(x, z_m) = \frac{Q}{U(z)} p_z(z_m - z_s, x - x_s), \tag{2.11}
\]

where subscript \(s\) denotes the source coordinates; \(p_z\) is the PDF of the particle height, which can be evaluated from the particle trajectories. In the following calculations \(z_s\) is equal to \(z_0\) and the time step is \(0.02\tau\), where \(z_0\) is the surface roughness length, and \(\tau\) is
the time scale \( \frac{w^2}{\varepsilon} \). In Eq. (2.5), \( <w(x,z_m)> \) is calculated by averaging the vertical velocities of the particles passing point \((x, z_m)\). Therefore, the footprint can be evaluated using Eqs. (2.4) and (2.5), which is also the basis of the derivation of the idealized model shown in a later section. Alternatively, given the distribution of the mixing ratio, the footprint can be found in the Eulerian field directly by integrating the advection-diffusion equation from the surface to height \( z_m \) (Horst and Weil, 1992), i.e.,

\[
Q^y(x,z_m) = -\frac{1}{Q} \frac{\partial}{\partial x} \int_0^{z_m} U(z) C^y(x,z) dz .
\] (2.12)

In deriving Eq. (2.12), the surface flux is \( Q\delta(x)\delta(y)\delta(z) \) at \((x, y, z) = (0, 0, 0)\), where \( \delta \) is Dirac Delta function, and the wind direction is assumed to be height-independent. The same results can be obtained using the two approaches.

**2.2.2 Results and discussion**

Two quantities, the cumulative footprint (CF) and footprint half-width, are introduced to characterize the footprint. The CF is defined as the horizontally-integrated footprint, i.e.,

\[
CF(X,z) = \int_0^X f^y(x,z) dx ,
\] (2.13)

where \( X \) is the dimensionless upwind distance \( (w_e x / Uh) \). The equilibrium CF value is referred to the CF value when \( X \to +\infty \). The footprint half-width is defined as the horizontal distance between the points where the footprint falls to one-half of its
maximum value (Weil and Horst, 1992). In the following sections, the CWIF, footprint half-width, and height are scaled by $w_*/U_h$, $U_h/w_*$, and $h$, respectively.

### 2.2.2.1 Case study

Figure 2.1 presents the cross-wind integrated footprint and the CF as a function of $X$ for different measurement heights in the case of $z_o = 10^{-3}h$ and $L = -0.03h$. The peak value of the footprint decreases with height, while the width increases with height. The area immediately upwind of the receptor ($X=0$), where the surface sources do not affect the measured flux (footprint is close to zero), becomes larger with height. The values of the footprint function are small at $X>1$ compared to those at $X<1$, indicating that the surface sources located at upwind dimensionless distance less than 1 contribute more to the measured flux. Another feature observed is that the footprint is slightly negative in the range $1<X<3$, as a result of large eddies in the CBL. Because of this feature, the CF increases monotonically from zero to its maximum value, and then decreases to its equilibrium value as $X$ increases (Figure 2.1b). This phenomena was described as ‘overshoot’ by Weil and Horst (1992).

Figure 2.2 (a-d) shows the variation of the peak value, the half-width, the peak location, and the equilibrium cumulative value of the footprint with height, respectively, under the same conditions as Figure 2.1. In the lower part of the CBL, the footprint peak decreases rapidly, and its location is farther away horizontally from the receptor with height. The half-width of the footprint increases with height in this part of the CBL. In
contrast, the half-width decreases with height in the upper part of the CBL due to the sizeable negative footprints. Therefore, the half-width defined above is not a suitable quantity to characterize the footprint function in that area. The equilibrium CF value decreases nearly linearly with height, i.e., \( CF(+\infty, z) = 1 - z/h \). This result is consistent with the linear flux profile in a well-mixed, horizontally homogeneous boundary layer when the entrainment flux is not considered (Wyngaard and Brost, 1984) because the equilibrium value \( CF(+\infty, z) \) can be interpreted as the flux at height \( z \) above the homogeneous surface with a unit surface flux (see Eq. 2.13) according to the footprint definition. This linear relationship also indicates that the flux signal becomes weaker as the measurement height increases.

### 2.2.2.2 Dependence of footprint on stability and roughness length

This section examines the dependence of the footprint for measurements in the mixed layer on atmospheric stability and surface roughness length.

Sixty-three experiments are carried out. The model is run for the cases of \( L = -0.01, -0.02, -0.03, -0.04, -0.05, -0.07 \), and \(-0.09 \) \( h \), in each of which nine experiments are made with \( z_\theta \) of \( 10^{-5}, 3 \times 10^{-5}, 5 \times 10^{-5}, 8 \times 10^{-5}, 10^{-4}, 3 \times 10^{-4}, 5 \times 10^{-4}, 1 \times 10^{-3}, \) and \( 2 \times 10^{-3} \) \( h \), respectively. The results from four cases are selected and shown in Figure 2.3. The width of the footprint broadens (Figure 2.3a), the peak value of the footprint decreases (Figure 2.3b), and the peak location of the footprint is farther from the source (Figure 2.3c) as the stability increases (larger \(|L|\)) because the vigor of vertical mixing
decreases. A change in the roughness length also affects the features of the footprint. Larger roughness usually indicates stronger mechanically-induced turbulent mixing, resulting in a narrower footprint half-width. In all cases, the equilibrium CF linearly decreases with height as shown in Figure 2.2d.

The features of the derived footprints can be fitted as functions of height, stability, and roughness length. The dimensionless half-width of the footprint, denoted as $\Delta X_{\text{stochastic}}$, can be described in the lower part of the CBL (less than $0.6h$) by a power-law profile,

$$
\Delta X_{\text{stochastic}} = g_1 \left( \frac{z}{h} \right)^{g_2},
$$

(2.14)

where coefficients $g_1$ and $g_2$ are polynomial functions of the stability parameter $L/h$ and the dimensionless roughness length $z_0/h$ (appendix A.2). The half-width in the upper part of the CBL is not fitted because the footprint is not well described by this quantity. In addition, direct measurements of the surface flux should not be made in the upper CBL in practice due to both the weaker surface flux signals and the impact of entrainment flux.

Similarly, the peak of the footprint, denoted as $F_{\max, \text{stochastic}}$, and its location ($X_{\max, \text{stochastic}}$) in the lower part of the CBL can be described by,

$$
F_{\max, \text{stochastic}} = g_3 \left( \frac{z}{h} \right)^{g_4} + g' \left( \frac{z}{h} \right)^3,
$$

(2.15)

and

$$
X_{\max, \text{stochastic}} = g_5 \left( \frac{z}{h} \right)^{g_6},
$$

(2.16)
where \( g_3, g_4, g_5, g_6, \) and \( g' \) are the coefficients determined by fitting the results from the stochastic model (see appendix A.2). It should be kept in mind that the parameters in Eqs. (2.14), (2.15), and (2.16) apply only when \( L \) and \( z_0 \) range from -0.01 to -0.1\( h \) and from \( 10^{-5} \) to \( 2 \times 10^{-3} h \), respectively. The fitted functions are shown by the solid lines in Figure 2.3.

These expressions are useful for making a quick estimate of the main features of the footprint given the CBL height, stability, and roughness length. Moreover, the streamwise distribution of the footprint is also needed to interpret flux measurements in detail. It is, however, inconvenient to apply the stochastic model to long term observations because it requires substantial computational resources. For the sake of simplification and computational expediency, an empirical analytical expression for the footprint function is proposed in the next section.

### 2.3 An approximate footprint model

This section starts with a simplified footprint model in an idealized CBL. The parameters in the simple model are then adjusted based on the results of the stochastic model.
2.3.1 An analytical footprint function under ideal conditions

To obtain an explicit analytical expression for the footprint above the surface layer, dispersion is considered in an idealized CBL with horizontally homogeneous conditions, uniformly-distributed horizontal wind speed, and vertical velocity skewness and variance in the vertical direction. In addition, the Lagrangian time scale is assumed to be infinite as adopted in the probability density function (PDF) dispersion models (Luhar, 2002; Misra, 1982; Weil et al., 1997). This assumption makes use of the observation that the Lagrangian time scale of the convective turbulence is so large that some passive particles tend to remain close to their initial trajectory for a considerable distance. Numerical simulations have justified the assumption indirectly. For example, Weil (1990) simulated the dispersion from a point source at the bottom of the CBL, and found that the large time scale is one of the important reasons for causing the phenomena that the plume rises or “lifts off” the surface as observed both in the laboratory experiments (Willis and Deardorff, 1976) and in field observation (Briggs, 1993a; Briggs, 1993b).

The above assumptions significantly simplify the solution of the vertical velocities and trajectories of particles (see Eqs. (2.6), (2.7), and (2.10)). Therefore, the initial vertical velocity of a particle passing the point \((x, z_m)\) is simply equal to,

\[
w_p(x, z_m) = \frac{(z_m - z_s)U}{x}, \tag{2.17}
\]

where \(z_s\) is the height of the receptor. Considering multiple perfect reflections on the top and bottom boundaries by applying Eq. (2.17) to virtual sources at \(z = \pm(2h-z_s), \pm(2h+z_s),\)
±(4h−z_s), etc., the vertical velocities of the particles passing (x, z_m) have the following initial values (Misra, 1982; Weil, 1988),

\[ w_{kj}(x, z_m) = \frac{(2kh + jz_m - z_s)U}{x}, \]  

(2.18)

where \( k \) is any integer whose absolute value is the number of times of reflections from the top of CBL; \( j \) is an integer equal to either 1 or -1, indicating reflection on the top or bottom boundaries. The cross-wind integrated mixing ratio contributed by the particles with the vertical velocity of \( w_{kj} \) is then (Misra, 1982; Weil, 1988)

\[ C_{kj}^r(x, z_m) = Q \frac{p_w(w_{kj})}{x}, \]  

(2.19)

where \( p_w \) is the probability density function of \( w_{kj} \) at the source height. It is usually taken to be the sum of two Gaussian distributions, which provides a good match to observations (Misra, 1982; Weil, 1988), i.e.,

\[ p_w(w) = \frac{\lambda_1}{\sqrt{2\pi}\sigma_1} \exp\left[ -\frac{(w - \overline{w}_1)^2}{2\sigma_1^2} \right] + \frac{\lambda_2}{\sqrt{2\pi}\sigma_2} \exp\left[ -\frac{(w - \overline{w}_2)^2}{2\sigma_2^2} \right], \]  

(2.20)

where \( \lambda_1, \lambda_2, \sigma_1, \sigma_2, \overline{w}_1, \overline{w}_2 \) are found by equating the zeroth through third moments of the assumed distribution with those specified and by assuming that \( \frac{\sigma_1}{\overline{w}_1} = -\frac{\sigma_2}{\overline{w}_2} = R \), a constant. Details on how to determine the six parameters can be found in the literature (e.g., Weil, 1990).

For the idealized case we are considering, the Eulerian vertical flux (Eq. 2.5) can be written as (appendix A.1),

\[ F_{m}^y (x, z_m) = \sum_{j=-1}^{+1} \sum_{k=-\infty}^{+\infty} w_{kj}(x, z_m) C_{kj}^r(x, z_m). \]  

(2.21)
Therefore, the combination of Eqs. (2.5), (2.19), (2.20), and (2.21) yields an expression of the footprint function according to Eq. (2.4),

\[
\begin{align*}
    f^x(x, z_m) = \begin{cases} 
        \frac{1}{X} \sum_{j=-\infty}^{\infty} \sum_{k=0}^{\infty} w_k j(x, z_m) p_n(w_k), & x > 0, \\
        0, & \text{otherwise}
    \end{cases}
\end{align*}
\]

which is actually a superposition of the functions (with different values of \(j\) and \(k\) in Eq. (2.22)) each of which includes a factor of \(\exp(-1/x^2)/x^2\). Assuming \(R=1\) and \(S=0.5\), typical values in the real CBL, we calculate the footprint as an example. Taking five reflections at the top boundary into account, i.e., a maximum of \(|k|=5\), can provide a sufficient accuracy for the calculation within \(X<4\).

Figure 2.4 shows the calculated footprint as a function of \(X\) for four measurement heights of 0.1, 0.3, 0.5, and 0.7\(h\). The footprint broadens with height, while its peak value decreases with height. Those results show general features of the footprint for the flux measurement in the CBL. The peak value, peak location, width, and equilibrium cumulative value of the footprint calculated from Eq. (2.22) vary with height (Figure 2.5), similar to those from the stochastic model (Figure 2.2).

However, the effects of stability, roughness, and variations in turbulence in the vertical direction are not included in the idealized model. These factors affect the acceleration of the particles and hence their trajectories, altering footprint simulations. Calculation of the footprint function requires numerical evaluation of the vertical velocities for more realistic wind, temperature, and turbulence profiles, and is not convenient for long-term calculations (e.g., interpreting multiple years of continuous flux...
measurements). Instead, we adjust Eq. (2.22) to approximate the more realistic solution of the footprint from the stochastic model.

2.3.2 Adjustment for the idealized model

Two parameters are introduced to adjust the main features of the footprint from the idealized model to those from the stochastic model. One, denoted by $\beta$, is used to adjust the predicted half-width and peak values of the footprint. The other, $\gamma$, is used to adjust the position of the maximum of the footprint by translating the footprint function a distance in the along-wind direction. A necessary condition that the equilibrium cumulative footprints of the adjusted and non-adjusted models are equal at the same height is imposed, i.e.,

$$
\int_{0}^{+\infty} f_{a}^{\gamma}(x, z_{m}) \, dx = \int_{0}^{+\infty} f^{\gamma}(x, z_{m}) \, dx = 1 - \frac{z_{m}}{h},
$$

where $f_{a}^{\gamma}(x, z_{m})$ is the adjusted footprint function which, therefore, can be written as,

$$
f_{a}^{\gamma}(x, z_{m}) = \begin{cases} 
\beta \frac{F_{m}(\beta x + \gamma, z_{m})}{Q} \\
\sum_{j=-\infty}^{\infty} \sum_{k=-\infty}^{\infty} \beta \frac{w_{j} \left( \beta x + \gamma, z_{m} \right) p_{a} \left[ w_{k} \left( \beta x + \gamma, z_{m} \right) \right]}{\beta x + \gamma},
\end{cases}
$$

for $\min[\beta x + \gamma, x] > 0$,

otherwise

where $\beta$ is a function of stability, roughness length, and height; it can be calculated by,
\[ \beta = \alpha \cdot \frac{\Delta X_{\text{ideal}}}{\Delta X_{\text{stochastic}}} + (1 - \alpha) \frac{F_{\text{max, stochastic}}}{F_{\text{max, ideal}}}, \]  
\hspace{1cm} (2.25)

where \( \alpha \) is a coefficient determining the relative adjustment weight. The parameter, \( \gamma \), is calculated by,

\[ \gamma = X_{\text{max, ideal}} - \beta X_{\text{max, stochastic}}, \]  
\hspace{1cm} (2.26)

where \( \Delta X_{\text{ideal}}, F_{\text{max, ideal}}, \) and \( X_{\text{max, ideal}} \) are the half-width, peak, and peak location of the CWIF as determined from the idealized model (see appendix A.2). Appendix A.3 shows the adjusted function Eq. (2.24) can satisfy the constraint Eq. (2.23) for a broad range of typical roughness lengths and atmospheric stabilities.

After adjustment, the analytical solution is generally in good agreement with that from the stochastic model. Figure 2.6 shows the comparison of the CWIFs obtained from the stochastic model and the adjusted model for two heights, \( z=0.21h \) and \( z=0.41h \), for \( L=-0.03h \) and \( z_0=10^{-3}h \). The coefficient, \( \alpha \), is taken as 0.5, indicating equal weights of the width and the peak of the footprint in the adjustment calculation. In the lower CBL, the result from the adjusted model is in good agreement with that from the stochastic model in the range of \( 0<X<1 \), which accounts for a majority of the footprint. For the higher altitude (0.41h), the adjusted model predicts a more rapid decrease in the footprint to the right of the peak than the stochastic model despite good agreement in the range \( X<0.65 \).

The performance of the adjusted model is poor far away from the receptor, i.e. \( X>1 \) or 2, due to oversimplified physics. The negative footprint and its location are not well simulated in the adjusted model, particularly under strongly unstable conditions or at high measurement levels. Nevertheless, the model still can be used to estimate the
footprint in the lower levels of the CBL because the portion of the footprint in the range of $X>1$ is small (~10%).

We compared the footprints estimated from the explicit model by Kljun et al. (2004) and the adjusted footprint model in this study for three receptor heights above the surface layer for $L$ values of -0.03$h$ and -0.07$h$ with $z_0=10^{-4}h$ (Figure 2.7). The predicted footprints from the two models are qualitatively consistent in terms of their dependences on the roughness length, stability, and the receptor height. The differences in the peak, peak location, and the width of the footprint are more significant at higher levels. Under strongly unstable condition, the model by Kljun et al. (2004) predicts footprints with smaller peaks, larger extensions, and peak locations farther away from the receptor than the adjusted model (Figure 2.7a). Similar differences are found for the two higher levels, and the differences are reversed at the lowest level under the less unstable conditions (Figure 2.7b). In addition, the model by Kljun et al. (2004) predicts small contributions of the footprints downwind of the receptor (i.e., $X<0$), which is not predicted in the adjusted model, and the adjusted model predicts negative footprints (for $X>1$), which is not predicted in the model by Kljun et al. (2004). The considerable differences are most likely due to the different stochastic models used. The parameters in the model by Kljun et al. (2004) are fitted by the results from a stochastic model with the effect of the longitudinal turbulence being considered, resulting in the larger extension and smaller peaks of the footprints (Kljun et al., 2003). Along-wind turbulence is not considered in the numerical model we used. Direct measurements are needed to assess the accuracy of the footprint simulations.
The footprints for measurements within the surface layer evaluated at the top of the surface layer (0.1\(h\)) using the model derived by Horst and Weil (1992; 1994) show higher peaks closer to the receptor and smaller extensions as compared to the other two models (Figure 2.7) because mean plume height is overestimated at high altitudes based on the surface layer turbulence (Weil and Horst, 1992).

**2.4 Application**

The adjusted footprint model is used to interpret CO\(_2\) fluxes measured at three levels on the 447-m WLEF tall tower over a mixed forested area in northern Wisconsin. One level is 30m, which is within the surface layer during the daytime. The other two levels are 122 and 396m, which are usually above the surface layer. Detailed descriptions of the flux measurements are given by Berger et al. (2001) and Davis et al. (2003). Upland forest and wetland (largely forested as well) vegetation dominate the land cover in the tower area (Figure 2.8). However, there is a grass-covered area of about 150-m radius centered at the tower base. The grassy area is included in the flux footprint area in some cases, possibly resulting in measurements unrepresentative of the predominant vegetation types.

As an example, we calculate the footprint during the day on June 4, 1998 (day 155) and overlay it on a vegetation cover map to assess the weightings of the flux from each vegetation type (Figure 2.8). The CBL is well developed in the midday hours on that day with fair weather. The M-O length is estimated using the meteorological variables and sensible heat fluxes measured at the tower. The depth of the mixed layer is obtained
from a 915-MHz boundary layer profiling radar (Yi et al., 2001). The roughness length and zero-plane displacement are estimated to be 0.5 m and 15 m, respectively, according to the observed wind profiles.

The flux footprint, \( f(x, y, z_m) \), can be written as the product of the cross-wind integrated footprint, \( f^y(x, z_m) \) and the crosswind mixing ratio distribution function, \( D_y(x, y) \) (Horst and Weil, 1992), i.e.,

\[
f(x, y, z_m) = f^y(x, z_m)D_y(x, y).
\]  

(2.27)

Dispersion in the lateral direction \((y)\) is assumed to be Gaussian, i.e.,

\[
D_y(x, y) = \frac{1}{\sqrt{2\pi\sigma_y}} \exp\left(-\frac{y^2}{2\sigma_y^2}\right),
\]  

(2.28)

where \(\sigma_y\) is the standard deviation of the plume in the lateral direction, and is usually formulated as a function of the atmospheric stability and the streamwise distance from the source via fitting observational data. To a first approximation, we use the formulation for \(\sigma_y\) based on both Passquill stability classes and the combination of the PG curves and results from experiments (Arya, 1999), i.e.,

\[
\sigma_y = c_1 x (1 + c_2 x)^{-1/2}
\]  

(2.29)

where \(c_1\) is equal to 0.32 for unstable conditions. The large roughness over this forested area is accounted for by setting the parameter \(c_2\) equal to 0.0004, which was designated originally for urban conditions (Arya, 1999). The model developed by Horst and Weil (1992; 1994) is used to estimate the crosswind integrated footprint within the surface...
layer, and the adjusted footprint model developed in this paper is used for the measurements above the surface layer.

Figure 2.9 shows the fractional contributions of fluxes from different vegetation types to the measured turbulent fluxes at the WLEF tower. These values are computed by integrating the footprint function over the areas covered by the corresponding vegetation. Footprint areas vary with atmospheric stability and wind direction, resulting in time-dependent contributions of each vegetation type to the total fluxes. At the 30m level, the fluxes from the grassy area at the base of the tower contribute about 20% of the measured fluxes depending on stability, while the flux sensors at 122m and 396m sample primarily from the areas covered by wetlands (forested and non-forested) and upland forest.

The mean source areas for the measurements at the three levels are presented in Figure 2.8. The three lines in the figure are the contours of footprints enclosing 80% of the total integrated footprint averaged from 9am to 5pm LST for the three levels. The source area of the lower level measurement is closer to the tower base, while the source areas at the higher levels are larger and do not include the grassy area at the tower base. The contributions of the wetlands and forest for the two high levels are similar and approximately independent of time in this case.

2.5 Discussion and Summary

The footprint for flux measurements above the surface layer under strongly unstable conditions has been studied using a Lagrangian stochastic model with state-of-the-art vertical turbulence parameterizations. The features of the footprint depend
strongly on stability, roughness length, receptor height, and CBL depth. The half-width and peak of the footprint increase and decrease with stability, respectively; the horizontal location of the peak is farther from the receptor with stability; the equilibrium cumulative footprint decreases linearly with height. The main features of the footprint at receptor heights below 60% of the CBL height can be described by simple polynomial functions, useful to quickly estimate the main features of the footprint in practice.

An analytical model for the footprint in an idealized CBL is derived. The idealized model reflects the overall characteristics of the footprint function predicted in the stochastic model. Due to the assumptions simplifying the model, the dependence of the footprint on stability and roughness length is not reflected in the simple model. The goal is to develop an explicit analytical but accurate footprint model that can be used to compute years of hour-to-hour flux footprints from tall tower measurements. To this end, the idealized footprint model is adjusted empirically to more closely match the main features of the stochastic model flux footprint above the surface layer and below 60% of the CBL depth. The solutions, both the position and streamwise extent of the footprint, from the adjusted model are in good agreement with those from the stochastic model for $X<1$, which covers the majority of the footprint. Applying a surface layer model to measurements above the surface layer in the CBL would result in significant errors in predicting the peak, peak location, and extent of the footprint.

It should be noted that there are limitations of the adjusted footprint model. Firstly, the model can be used only below 60% of the CBL depth because the fit is only accurate below $0.6h$. In addition, the effect of entrainment on the footprint, more significant with increasing height, is not considered in either the stochastic or the
idealized footprint models. In reality, measuring surface fluxes far away from the surface yields very little information about the surface flux. Secondly, the model applies to stabilities of $0.01 < -L/h < 0.1$ and roughness lengths of $10^{-5} < z_0/h < 2 \times 10^{-3}$, where the model is empirically adjusted. Under extremely unstable conditions, the effects of the along-wind turbulence would be significant, and is not considered in our model. Thirdly, researchers should avoid using the adjusted model close to the surface ($z_m < 0.05h$). The main features of the footprint are better represented by the fitted curves above the surface layer than within the surface layer, for few data points below the surface layer are fitted. Finally, the model applies only over a dynamically homogeneous surface like other analytical footprint models in the literature.

The negative footprint, which is more significant with height and found at a distance of approximately $X > 1$ or 2, is not well simulated by the adjusted model. Because only a small portion of the total footprint is contributed by the range $X > 2$, the adjusted model still accurately simulates the main features of the footprint above the surface layer. The present model can be improved by considering more physics and be better adjusted based on direct footprint measurements or simulations from more realistic models.

Comparison of the footprints simulated by the adjusted model in this study and the model proposed by Kljun et al. (2004) shows similar dependence of the footprint on stability, roughness length, and receptor height, but significant differences quantitatively in the peak value, peak location, and streamwise extent of the footprint particularly near the middle of the CBL. The differences are probably due mainly to the different performances of the stochastic models that are used to determine the parameters in the
empirical models. Measurements are needed to make further assessments about the accuracy of the existing models and to improve the ability of footprint simulations.

Despite the limitations, the analytical solution renders the adjusted footprint model very convenient to use in practice. The model is used to estimate the sampling areas of flux measurements at 122m and 396m above the ground at the 447-m tall WLEF tower. The footprints are compared to those for the measurements within the surface layer (30m). Under strongly unstable conditions, the lowest level flux is found to have significant flux contributions from the grassy area surrounding the tower base. The case study indicates that the sampling areas of the top two levels in the day consist mainly of wetland and upland forest areas while the lowest level measurements sample areas of grass, wetland, and upland forest; hence daytime measurements at the top two levels might better represent the daytime wetland and upland forest fluxes in this region.

Acknowledgement:

We thank Dr. A. Luhar who provided the source code of the Lagrangian stochastic model used in this study.
Figure 2.1: Normalized (a) footprint (CWIF) and (b) cumulative footprint derived from the stochastic model as a function of the horizontal dimensionless distance ($X$) from the receptor ($X=0$) for observation heights of 0.11, 0.31, 0.51, and 0.71$h$ in the case of $L/h=-0.03$ and $z_0/h=10^{-3}$. Wind direction is from right to left.
Figure 2.2: Normalized (a) peak, (b) peak’s location, (c) half-width of the footprint, and (d) equilibrium cumulative footprint as a function of dimensionless height \( z/h \) in the case of \( L/h = -0.03 \) and \( z_0/h = 10^{-3} \).
Figure 2.3: Dependence of footprint (CWIF)’s (a) half-width, (b) peak, and (c) peak location on the stability and roughness length as a function of height. The solid lines are the fitted lines for the three quantities. The three quantities are normalized (see text).
Figure 2.4: Normalized cross-wind integrated flux footprint, $f^y$, calculated from the ideal model (Eq. 2.22) with a vertical velocity skewness of 0.5 and $R=1$, as a function of the horizontal dimensionless distance ($X$) for the heights of 0.1, 0.3, 0.5, and 0.7$h$. 
Figure 2.5: Normalized (a) peak, (b) peak’s location, (c) half-width of CWIF, and (d) equilibrium cumulative footprint derived from the idealized model as a function of dimensionless height ($z/h$).
Figure 2.6: Comparison of the footprints (CWIF) calculated from the adjusted model and from the stochastic model for heights of 0.21 and 0.41h in the case of $L/h = -0.03$ and a typical roughness length for forest, $1 \times 10^{-3} h$.
Figure 2.7: Comparison of the predicted cross-wind integrated footprints from the adjusted model of this study (Eq. 2.24) (solid lines) and from the empirical model (dotted lines) by Kljun et al. (2004) at three heights (colors) above the surface layer with $L=-0.03h$ (a) and $L=-0.07h$ (b) for $z_0=10^{-4}h$. The long dashed lines are the results at the top of the surface layer ($0.1h$) from the surface layer model by Horst and Weil (1994).
Figure 2.8: Vegetation map centered at the WLEF tower. The cross represents the location of the tower. The data is from the state of Wisconsin’s department of Natural Resources, http://www.dnr.state.wi.us/maps/gis/datalandcover.html. The three lines represent the 80% source area boundaries (see text), calculated using the averaged footprint from 0900 to 1700 on June 4, 1998, for the measurement heights of 30, 122, and 396m, respectively.
Figure 2.9: Fractional weights (in colors) by area of the flux footprint from wetland, forest, grass, and other vegetation to the total turbulent flux measured at the heights of (a) 30, (b) 122, and (c) 396m above the ground level as a function of time on June 4, 1998. Note that the footprints for the measurements at 30m are computed with the roughness length and zero-plane displacement being equal to 0.2m and 6m, respectively, to roughly consider the effects of the grassy area.
3.1 Introduction

Terrestrial ecosystems play a critical role in the global carbon cycle. To date, there is still significant uncertainty about how much carbon dioxide (CO₂) is absorbed by terrestrial vegetation and what factors control this process (IPCC, 2001), leading to large uncertainty in predictions for the future uptake or release of CO₂ in the global environment (Cao and Woodward, 1998; Cramer et al., 2001; Huntingford et al., 2000; IPCC, 2001; Pan et al., 1998; Woodward and Kelly, 1995; Woodward et al., 2002). To understand fully the role of terrestrial ecosystems in the global carbon cycle, we need to test mechanistic understanding of the carbon balance gained at smaller scales against observations obtained at regional and global scales. This requires observations of the exchange of CO₂ between the atmosphere and terrestrial ecosystems over various scales, e.g., from a forest stand area (about 1 km²) to the entire globe. Currently, the net ecosystem-atmosphere exchange of CO₂ (NEE) can be inferred on global and somewhat on continental scales by means of inverse modeling (Bousquet et al., 1999a; Ciais et al.,
1995; Enting et al., 1995; Gurney et al., 2002; Tans et al., 1990), and NEE can also be
directly observed on local scales (areas of order 1 km$^2$ or smaller) using techniques such
as tower-based eddy covariance (Baldocchi et al., 2001). The exchange of CO$_2$, however,
is not readily measured for regions between 1 km$^2$ and the globe in area, limiting our
understanding of the mechanisms governing the global carbon cycle.

It is more difficult to extend local measurements across a region with a
heterogeneous ecosystem than a region with a homogeneous ecosystem because tower- or
aircraft-based flux measurements over a mixed land cover area either may not be
representative of the fractional coverage of land cover types within the region, or may
miss some important cover types. One approach to estimating CO$_2$ fluxes in such a region
is to extrapolate flux measurements encompassing a subset of the domain using a
landscape classification scheme, assuming that the flux measurements capture a
representative sample of ecosystem types in the region. Regional fluxes can then be
estimated from the flux measurements and proportions of the identified ecosystem types
that constitute the region. This aggregation is straightforward, but difficult to carry out in
practice due to the demands of identifying important ecosystem types, and then gathering
sufficient information on both the distribution of the ecosystems of all types and the
fluxes for each ecosystem. An assumption of similarity of ecosystems within a landscape
is needed for spatial aggregation because finely resolved classification data, e.g. at the
species level, is not available in practice (Mackay et al., 2002). No general theory exists
to determine what level of ecosystem classification is acceptable in terms of both the
accuracy of the classification data and the representativeness of the measurements for
each ecosystem type. In principle, regional fluxes can be estimated more accurately by conducting measurements for more ecosystem types.

It is, however, usually impractical to make flux measurements for a large number of ecosystem types in a region. Alternatively, fluxes for multiple ecosystem types can be inferred by decomposing the measured fluxes over mixed ecosystems, in which eddy-covariance fluxes are interpreted as the weighted averages of the fluxes for all ecosystem types in their footprint areas (Horst and Weil, 1994). Assigning eddy-covariance fluxes to multiple ecosystems in the footprint area also benefits the interpretation of the measurements. Chen et al. (1999), for example, proposed a scheme to separate aircraft flux measurements into fluxes for specific land cover types by solving a set of linear equations. In their approach, a flux record is separated into a number of segments, each of which is regarded as a weighted average flux of all land cover types in the corresponding footprint area. If the number of segments is equal to or more than that of the land cover types, the fluxes of each land cover type can be solved. Ogumjemiyo et al. (2003) used a multiple regression model to relate aircraft flux measurements to the fractional distribution of surface land cover types. Both studies indicate that conducting flux decomposition is a promising approach to estimate the NEEs for specific ecosystem types. No similar analyses, however, have yet been applied to tower-based measurements, whose long duration in time can capture long-term flux variability and integrals.

In northern Wisconsin NEE measurements have been made at three levels on a 447m-tall tower over a mosaic of upland and wetland ecosystems for several years (Davis et al., 2003; Ricciuto et al., submitted). We attempt to infer the fluxes for the dominant
ecosystems around the tall tower by decomposing the measured fluxes using an eddy-covariance flux footprint model. Demonstrating this approach is the first goal of this paper. The second goal is to identify, using the flux decomposition, the key ecosystem types in the region around the tall tower. The third goal is to examine the aggregated fluxes in a region using ecosystem characteristics inferred from the decomposed fluxes. To achieve those goals, a method has been developed to derive the stand-level NEE for the typical types of the ecosystems in the region from the tall tower observation. In addition the region around the tall tower includes a large number of stand-level eddy-covariance flux measurements, conducted by a variety of investigators (e.g., Desai et al., 2005; Desai et al., submitted; Noormets et al., submitted). In particular, NEE has been measured for several years over two nearby stands, a lowland shrub wetland and a mature deciduous upland (Cook et al., submitted; Cook et al., 2004). In an effort to interpret the signal from the tall tower, we use these two towers as an alternative regional up-scaling approach, and as an attempt to evaluate the flux decomposition. Desai et al. (submitted) does a more extensive flux tower upscaling using the full suite of eddy-covariance measurements (eleven different stands, not including the tall tower) available in the region. Section 3.2 describes the details of the method and data used. A stand-level classification scheme is introduced as the framework for making comparisons and aggregating fluxes to the region. Results and comparisons are presented in section 3.3. Discussion is presented in section 3.4. The aggregated regional fluxes are discussed in section 3.5. Section 3.6 discusses limitations of our results and future work.
3.2 Materials and methods

3.2.1 Site and data used

3.2.1.1 Site description

The study site is located in the Chequamegon National Forest, centered at a tall communication tower (WLEF) (45.9455878° N, 90.272304°W). The tower is about 15 km east of Park Falls, Wisconsin. A 60km×60km land cover map is presented in Figure 3.1. The spatial resolution of the map is 30m. The land cover data provided by WISCLAND (Wisconsin Initiative for Statewide Cooperation on Landscape Analysis and Data) were derived from LANDSAT Thematic Mapper (TM) satellite imagery (WiDNR, 1998). Dominant vegetation types in this region are mixed coniferous and deciduous forests, lowland and wetland forest. The deciduous species include aspen, birch, maple, basswood and alder. Deciduous forests comprise about 70% of the landscape according to the WISCLAND classification scheme (WiDNR, 1998). The coniferous species include balsam fir, red pine, jack pine, black spruce, and white cedar. Topography in the region is flat to gently sloping. More detailed descriptions of vegetation in the vicinity of the WLEF tower can be found in Mackay et al. (2002) and Burrowes et al. (2002).
3.2.1.2 Tower-based data used

The NEEs of energy, water, and CO₂ are measured using the eddy-covariance method at the WLEF, Willow Creek (WC), and Lost Creek (LC) towers, whose locations are shown in Figure 3.1. Flux measurements have been conducted at three levels, i.e., 30, 122, and 396m, on the 447-m WLEF tower since 1995 (Davis et al., 2003). The height of the WLEF tower and multiple levels of flux instrumentation are unique within the current global flux tower network. CO₂ mixing ratio data, traceable to WMO primary standards, are collected at 11, 30, 76, 122, 244, 396 m (Bakwin et al., 1998). Air temperature, humidity, wind speed and direction are measured at 30, 122, and 396m. Other micrometeorological variables, e.g., radiation, soil temperature and moisture profiles etc., are measured at several locations in the region around the tower. Canopy height, while highly variable from stand to stand, reaches a maximum of 20-25m.

At the WC tower, flux data are collected at 30m above the ground. CO₂ mixing ratio, wind, temperature and humidity data are collected within and above the canopy, which is about 20-25m tall. Radiation, soil temperature, and soil moisture profiles are measured. The tower is located at a mature upland forested area with mixed deciduous species dominated by sugar maple, with some basswood and aspen. Detailed descriptions of the instrumentation and land cover at this site are given by Cook et al. (2004). Similar to measurements at WC tower, high-frequency flux and micrometeorological profile data are collected at LC tower except that the LC tower is 10m high. The tower is located at a mixed lowland and forested wetland area. Canopy height is typically 1-2 m and consists primarily of alder and willow.
Similar measurement methodology and data processing techniques are applied at the three towers, as documented by Berger et al. (2001), Davis et al. (2003), and Cook et al. (2004). Additional details about the sites, measurements and research activities can be found at http://cheas.psu.edu. Vegetation covers typically within the daytime flux footprints are mixed wetland and upland forests at WLEF (122 and 396m levels), upland deciduous forests at WC, and wetland forests at LC. Footprint modeling suggests that under unstable atmospheric conditions the 30m level of the WLEF tower is significantly influenced by the grassland around the tower base.

3.2.1.3 Ecosystem classification

In this study, ecosystems are classified based on plant function and watershed function, following categories that have been shown significant in a study of regional evapotranspiration (Mackay et al., 2002). Although a detailed classification of land covers is provided in the WiscLAND data product (WiDNR, 1998), we regroup them for two major reasons. First, the system of equations to be solved becomes intractable if there are too many unknowns, and the number of unknowns is proportional to the number of land cover types. Therefore, any land cover type whose percentage area within the footprints for flux measurements is smaller than 5% is combined with others. Second, significant differences in characteristics of NEE within the types of upland forests and within the types of wetlands are taken into account, which potentially alters the results of landscape-scale water and CO₂ fluxes estimated by the aggregation approach. Mackay et al. (2002) examined how forest species types in northern Wisconsin affect landscape...
scale water fluxes predicted from models driven by remotely sensed forest classification and sap flux data, pointing out that the distinction between aspen and other hardwoods is needed because of high growth rate of the aspen. In consideration of the above factors, the ecosystems are classified into six types as shown in Table 3.1. The ecosystems are also classified into wetland and upland according to watershed functions as a comparison. The fractional areas of the six stand types in a 40-km square region centered at the WLEF tower are also listed in Table 3.1. In the calculation, CO₂ fluxes for open water and roads are assumed to be zero.

3.2.2 Equations and methods

The vertical turbulent flux measured at height $z_m$ can be related to the spatial distribution of surface fluxes through a footprint function (Horst and Weil, 1994; Schuepp et al., 1990), i.e.,

$$ F_m(x, y, z_m) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} F_0(x', y', 0) f(x - x', y - y', z_m) dx' dy', \quad (3.1) $$

where $x$ and $y$ are the horizontal coordinates; $F_0$ and $F_m$ are the fluxes at the surface and measured at $z_m$, respectively; $f$ is the footprint function describing the contribution of each unit element of the upwind surface area to $F_m$.

Assuming that the whole ecosystem can be classified into $n$ types and the NEE for a specific ecosystem type is independent of location, we can rewrite Eq. (3.1) for the measurement at $(0, 0, z_m)$ when $z_m$ is within the surface layer as,
where \((\text{NEE})_i\) is the NEE for the \(i\)th type of ecosystem, which is unknown and to be determined; \(w_i\) denotes the weight of the NEE from the ecosystem type \(i\) to the measured flux. When the measurement height is above the surface layer under the convective condition in the day, Eq. (3.2) can be rewritten as,

\[
(1 - \frac{z_m}{h})\text{NEE}_m(0,0,z_m) \approx \sum_{i=1}^{n} [(\text{NEE})_i \times w_i],
\]

where \(h\) is the convective boundary layer (CBL) height; \(\text{NEE}_m\) is the measured surface flux. In deriving Eq. (3.3), we have assumed that the vertical turbulent flux varies linearly with height from the surface to the CBL top (Wyngaard and Brost, 1984). In this case, the term on the left hand side of Eq. (3.3) is an approximation to the contribution of the surface flux to the measured vertical flux, which can be described by Eq. (3.1) because the footprint function is derived under the assumption of zero entrainment flux. The weight \(w_i\) can be expressed as,

\[
w_i = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} H(x',y')f(-x',-y',z_m)dx'dy',
\]

where \(H\) is a sign function that is equal to 1 if the ecosystem at location \((x', y')\) belongs to type \(i\), and 0 otherwise. Given the footprint function and the distribution of ecosystems (and hence \(w_i\)), it is theoretically possible to estimate NEEs for the \(n\) ecosystem types \((n=6\) in this study, see Table 3.1) by solving a set of linear equations as long as the number of the flux measurements for a given time is equal to or greater than the number of the ecosystem types. In practice, flux records from aircraft measurements can be divided into several segments equal to or more than the number of ecosystem types. The
fluxes on the segments of a flight leg can be treated as simultaneous measurements at different locations under the assumption of negligible temporal variation in the fluxes. With footprint models and a vegetation map, the fluxes over a mixed land cover area can be decomposed into NEEs by ecosystem types (Chen et al., 1999; Ogunjemiyo et al., 2003). We call this approach the direct decomposition of NEE in this paper, for the NEEs can be directly determined. For tower-based measurements, direct decomposition requires fluxes measured simultaneously in the region at several towers or heights, or both, whose total number is at least equal to the number of the ecosystem types. It is, however, impractical to solve for the NEE of a large number of ecosystem types without flux measurements at a large number of towers or heights. Alternatively, we can take advantage of the long temporal record of continuous tower-based flux measurements. The footprints of the measurements vary with meteorological conditions such as wind direction and stability, leading to time-dependent sampling areas and $w_i$. Using ecosystem models to express the NEE of CO$_2$ as functions of environmental variables, we can solve for the parameters in the models for different ecosystems using Eq. (3.2). We call this approach indirect decomposition, for the NEE values are calculated using parameters derived in the ecosystem models. This decomposition can span a long period of time.

### 3.2.2.1 Ecosystem models

A widely-used rectangular hyperbolic equation (Ruimy et al., 1995) is selected to describe daytime NEE response to light intensity,
where \( I \) is the photosynthetically active radiation (PAR), \( \alpha \) is the apparent quantum yield (the slope of the light curve at \( I=0 \)); \( P_m \) is the maximum assimilation rate at saturated PAR, and \( R_d \) is the dark respiration rate (i.e., NEE at \( I=0 \)) that is assumed to be constant in this model.

The ecosystem nighttime respiration rate, \( R \), is usually described as a function of temperature. A review of respiration models is given by Lloyd and Taylor (1994). In this study, the Van’t Hoff’s exponential model is selected to describe the ecosystem respiration rate at night, which is mathematically equivalent to the widely used \( Q_{10} \) equation,

\[
R = R_{10} Q_{10}^{\frac{T-10}{10}}, \tag{3.6}
\]

where \( T \) is soil or air temperature in degrees C, \( R_{10} \) is the reference respiration rate at \( T=10^\circ \text{C} \); \( Q_{10} \) represents the increase in respiration for every 10 degree rise in temperature. Although more complicated models, e.g., considering the effects of moisture, are needed in some cases (Reichstein et al., 2002; Ricciuto et al., submitted), the exponential model is selected for its simplicity to test our decomposition method, as is the light response model Eq. (3.5). Simple models, similar to a limited number of ecosystem types, improve the stability of the inversion calculation. The air temperature measured at 30m on the WLEF tower is used in the calculation in the expectation that it is better representative of temperature in the broad, mixed forest footprint area than the soil temperature measured at any single point.
3.2.2.2 Footprint models

The flux footprint, \( f(x, y, z_m) \), can be written as the product of a crosswind-integrated footprint, \( f^y(x, z_m) \), and a crosswind concentration distribution function, \( D_y(x, y) \),

\[
f(x, y, z_m) = f^y(x, z_m) \times D_y(x, y), \tag{3.7}
\]

where \( x \) is the upwind distance and \( y \) is the crosswind distance from the centerline. Dispersion in the crosswind direction is assumed to be symmetric. \( D_y \) is modeled as a Gaussian function,

\[
D_y(x, y) = \frac{1}{\sqrt{2\pi\sigma_y(x)}} e^{-\frac{y^2}{2\sigma_y(x)^2}}, \tag{3.8}
\]

where \( \sigma_y \) is the standard deviation of the plume in the crosswind direction, which is usually formulated as a function of the atmospheric stability and the streamwise distance from the source via fitting observational data. We use a formulation for \( \sigma_y \) from Arya (1999) which uses Pasquill stability classes and empirical results from field measurements,

\[
\sigma_y = c_1 x (1 + c_2 x)^{-1/2}, \tag{3.9}
\]

where \( c_1 \) is equal to 0.32 under unstable conditions and 0.16 under near-neutral conditions. The parameter \( c_2 \) is selected as 0.0004 to take into account relatively the large roughness values over the forest.

The crosswind integrated footprint function, \( f^y \), is estimated within the surface layer using the model derived by Horst and Weil (1994; 1995), which combines solutions from analytical and Lagrangian stochastic dispersion models for the cross-wind
integrated concentration distribution for near-surface sources (Horst, 1979; van Ulden, 1978). The footprint can be written as (Horst and Weil, 1994),

\[ f^y(x, z_m) = \frac{\Phi}{z_m} \frac{d\bar{z}}{dx}, \]  

(3.10)

and,

\[ \frac{d\bar{z}}{dx} = \frac{k^2}{[\ln(p\bar{z}/z_0) - \psi_m(p\bar{z}/L)] \phi_h(p\bar{z}/L)}, \]  

(3.11)

where \( \bar{z} \) is the mean plume height for dispersion from a surface source, \( p=1.55, k=0.4 \) is the von Karman constant, \( z_0 \) is the surface roughness length, \( L \) is the Monin-Obukhov length, and \( \phi_h \) and \( \psi_m \) are the M-O stability correction functions (Businger et al., 1971).

The dimensionless function \( \Phi \) is, approximately,

\[ \Phi \approx \left( \frac{z_m}{\bar{z}} \right)^2 \frac{\overline{\eta}(z_m)}{U(\bar{z})} \frac{A e^{-\left(\frac{z_m}{\bar{z}}\right)}}{r}, \]  

(3.12)

where \( \overline{\eta}(z) \) is the mean wind speed profile, \( U \) is the mean velocity of plume advection, \( r \) is an empirical parameter which ranges from 1 to 2 depending upon stability, \( A=r \Gamma(2/r) / \Gamma^2(1/r) \), and \( b=\Gamma(1/r) / \Gamma(2/r) \). The parameter \( r \) is assumed to be 1.5 in these calculations. Details concerning the surface layer footprint model can be found in Horst and Weil (1994).

The empirical formula described in Chapter 2 is used to estimate the crosswind-integrated footprint for measurements above the surface layer under strongly unstable conditions. An analytical solution for an idealized CBL is adjusted to more closely match a stochastic model with more realistic atmospheric conditions (Wang et al., 2004).
In this study, the Monin-Obukhov length is calculated using the sensible heat and momentum fluxes directly measured at the tall tower. The roughness length, $z_0$, and zero-plane displacement, $d$, are estimated using logarithmic wind profiles for near-neutral conditions. $z_0$ and $d$ are estimated to be in the range of 0.8 to 1.1m and 15.5 to 16.4m, respectively, using the measured wind speeds at 30 and 122m. Therefore, $z_0$ and $d$ are approximated as 0.9 and 16m, respectively, for computing the footprints of the measurements in the CBL within which land covers are mostly mixed forested wetlands and uplands. CO$_2$ is assumed released or absorbed at the level of $d$ above the surface when footprints are estimated. However, applying the values of the parameters to computing the footprint for the measurements within the surface layer gives contradictory results, due to the uncertainty of the footprint simulation. As a first order approximation, we take $z_0$ and $d$ as 0.2 and 6 m, respectively, for computing the surface layer footprints in the following calculations. The reason for picking these values and the sensitivity of results to the surface layer footprint uncertainty will be discussed later (section 3.3.4). More advanced models are needed to simulate footprints robustly in the future study. CBL height is estimated using the empirical formula derived from radar data during campaigns in 1998 and 1999 at the WLEF tower (Yi et al., 2001).

### 3.2.2.3 Solving the equations

Substituting Eqs. (3.5) and (3.6) into Eq. (3.2), we can write the measured daytime fluxes as,
and at night as,

\[ F_m(t) = \sum_{i=1}^{n} \left[ \frac{\alpha_i P_{mi}(t) I(t)}{\alpha_i I(t) + P_{mi}} + R_{d_i} \right] \times w_i(t), \quad (3.14) \]

where \( \alpha, R_d, P_m, R_{10}, \) and \( Q_{10} \) with the subscript \( i \) represent the parameters defined in the ecosystem flux models for the \( i \)th ecosystem type; \( I, T, w_i \) and \( F_m \) are functions of time \( t \).

The time series of the measured fluxes and estimated weights comprise a set of equations with \( 3 \times n \) and \( 2 \times n \) unknowns for day and night data, respectively. As long as the number of flux measurements is greater than or equal to the number of unknowns, the unknowns can be determined. In this study, we used the Levenberg-Marquardt algorithm (Rodgers, 2000), which combines the steepest descent and inverse-Hessian function fitting methods, to find the optimal solutions for the nonlinear equations by minimizing the following objective function,

\[ J = \sum_{j=1}^{N} \frac{(y_j - Y_j(\hat{p}))^2}{\sigma_j^2} + \beta \sum_{k=1}^{N_p} \frac{(P_k - P_{\hat{k}})^2}{\sigma_k^2}, \quad (3.15) \]

where \( N \) is the total number of the data points; \( N_p \) is the number of the parameters unknowns of the model; \( y_j \) is the measured flux; \( Y_j \) is the modeled flux with the \( N_p \) parameters (\( \hat{p} : p_1, p_2, \ldots, p_k \)); \( P_k \) is the prior information for the \( k \)-th parameter, which is estimated by fitting Eq. (3.5) or Eq. (3.6) to WLEF data and served as the initial value; \( \sigma_j \) and \( \sigma_j' \) are the standard deviations of \( y_j \) and \( P_k \). The second term on the right side of the above equation can be interpreted as the extended-data used to constrain the solution (Enting, 2002). Since we are uncertain about how to quantify errors in the measured
fluxes and the simulated footprints, the data are unweighted, i.e., the weights for the data
used are uniform, in the following experiments. The non-negative parameter $\beta$ is
introduced to adjust the weight of the prior information. The solutions are in general
closer to the prior values as the value of $\beta$ increases. But not many stable solutions can be
obtained if $\beta$ is too small for the cases at night. Numerical experiments suggest that the
solutions can distinguish the parameters among the ecosystem types with $\beta$ of 0.1(day)
and 1(night). The central values and uncertainties for the solved parameters are estimated
as the arithmetic mean and the standard deviation of the results for an analyzed period
such as a month. This calculation is performed on blocks of data that are seven days to a
few weeks long, depending on the availability of the data.

3.3 Experimental Results

Flux and micrometeorological data collected in 2000 and 2003 are used to
estimate the ecosystem model parameters. Data in 2001 are excluded because a tent-
caterpillar outbreak significantly altered summertime fluxes (Cook, 2004). Flux data at
the WLEF tower in 2002 are not reliable because of an error in the data collection
system. Results based on WLEF and WC data for 2000 and 2003, and LC data for 2003
(the tower was installed in the midst of the 2000 growing season) are combined to
conduct statistical tests. In the following analyses, the growing season is defined as the
months of May through September. Unpaired and paired $t$-test methods are used for
comparing the results between sites, and among the ecosystems in the WLEF area,
respectively. Parameters are considered significantly different when $p$-values are smaller than 0.05, and slightly different when $p$-values are between 0.05 and 0.32, where $p$ is the probability that the null hypothesis was true (Ott, 1993).

### 3.3.1 Daytime parameters for different ecosystems

The parameters are estimated using the hourly-mean data grouped in windows of seven consecutive days. The window is sequentially marched through the data record by increments of one day for the time period from May through September in each of the two years. For any period of data, all available daytime data (PAR > 0) with neutral to unstable conditions (buoyancy flux > 0) within the surface layer (estimated as 10% of the CBL height) are adopted. For measurements above the surface layer, only data from strongly unstable conditions ($|L|$ is smaller than 10% of the CBL height) are retained because the footprints under other stability conditions cannot be well quantified. Solutions are rejected if they exceed the range of reasonable values from the literature: $-1 < \alpha < 0$, $0 > P_m > -100 \, \mu\text{mol m}^{-2}\text{s}^{-1}$, and $0 < R < 15 \, \mu\text{mol m}^{-2}\text{s}^{-1}$ (Ruimy et al., 1995).

#### 3.3.1.1 Seasonal patterns

Seasonal patterns of all ecosystem model parameters derived from data collected during the growing season of 2000 are shown in Figure 3.2. Light-saturated assimilation and daytime respiration rate, two parameters in the light response model, peak generally
in summer for all ecosystems. The patterns are similar to those derived from the WC flux measurements. In contrast to those parameters, not all the apparent quantum yields, $\alpha$, of the ecosystems vary significantly with season, which is consistent with what is reported in many other studies (e.g., Luo et al., 2000; Nilsen and Sharifi, 1994).

### 3.3.1.2 Seasonal average quantum yield, $\alpha$

To quantitatively compare the parameters among the ecosystems, Table 3.2 shows their average values and standard deviation from June through August of 2000 and 2003. (In this case, results in May and September are not included to reduce the seasonal variability in the comparisons of the parameters among ecosystem types.) No significant differences in the quantum yields estimated for the ecosystems (types I- VI) in the WLEF footprint area and observed at the tower sites are found. The overall value of the average quantum yield is about -0.05 $\mu$molC/$\mu$mol quanta.

### 3.3.1.3 Seasonal average light-saturated assimilation rate, $P_{\text{max}}$

In the WLEF footprint area, the forested wetland ecosystem (type V) and aspen ecosystem have larger magnitudes of light-saturated assimilation rates ($P_{\text{max}}$) than the others (Table 3.2). The aspen ecosystem has the largest $P_{\text{max}}$ among the three upland forest ecosystems (types I, II, and III) in the WLEF area, implying (1) strongest potential for CO$_2$ uptake in the day, and (2) different responses of the various upland ecosystems to
the same environmental conditions. The second implication is significant for aggregating fluxes from stand levels to regions.

The magnitude of $P_{\text{max}}$ of the aspen ecosystem (type II) is significantly larger than that of the mature upland deciduous forests (type III), consistent with the high growth rate of aspen relative to later successional species such as maple and supporting the need to distinguish between aspen and later successional deciduous forests for flux aggregation in this region as reported in a previous study of evapotranspiration using sap flux measurements (Mackay et al., 2002). The light-saturated assimilation rate measured at the WC upland deciduous site is close to that inferred for the aspen ecosystem in the WLEF area, but larger than that for the mature upland deciduous forests (type III).

With respect to the wetland ecosystems, the forested wetland ecosystem has a larger magnitude of $P_{\text{max}}$ value than the lowland shrub wetland ecosystem. The light-saturated assimilation rate measured at the LC site (most similar to the lowland shrub wetland ecosystem) is smaller than that of the same ecosystem type in the WLEF area. The “other” category, a mixture of shrub, grassland, agriculture, etc., has a small value of $|P_{\text{max}}|$ compared to forested ecosystems. This category is primarily represented by the grassy clearing around the WLEF tower which can be detected by the 30m flux measurements under unstable conditions.

3.3.1.4 Seasonal average daytime ecosystem respiration rate, $R_d$

The intercept of the light response curve when PAR is equal to zero, i.e., $R_d$ in Eq. (3.5) can be interpreted as the mean respiration rate of an ecosystem in the day,
which is one of methods used to estimate the respiration rate despite uncertainty (Ruimy et al., 1995). Among the three measurement sites, LC and WLEF sites have the smallest and largest seasonal-average respiration rates, respectively (Table 3.2). The respiration rate at the WC site is in between. The respiration rates estimated for all three types of upland forest ecosystem (types I, II, and III) in the WLEF footprint area are larger than the rate derived using data measured at the WC upland deciduous forest site. Similarly, the respiration rates for both types of wetland ecosystem (types IV and V) in the WLEF area are significantly larger than that for the wetland ecosystem at the LC site, respectively. Those comparisons imply that the respiration rates measured at WC and LC differ in some way from the corresponding type of ecosystem in the WLEF footprint. This presents serious difficulty for regional upscaling according to the land-cover-based ecosystem classifications.

Among the six types of ecosystem in the WLEF footprint area, not all the inverted results show statistically significant differences in $R_d$. Because of the critical role of respiration in the regional carbon cycle, the respiration rates are further analyzed using nighttime data in the next section, which may be more robust.

### 3.3.2 Nighttime functional parameters for different ecosystems

Nighttime data (PAR =0) with near-neutral conditions, e.g., $|L| >300$ m, are selected to estimate the parameters in the respiration model for the six ecosystem types. In this case, only measurements at 30m are selected, and the surface layer footprint model is employed to calculate the weights in Eq. (3.4). The data are limited to near-neutral
conditions because current footprint models are not well-defined in the stable atmosphere. In practice, the flux measurement is often decoupled from the surface under very stable conditions. The selection criteria result in many nighttime data points being screened out. As a result, the data window width is expanded to 20 days to ensure that a sufficient number of data points are available to solve the set of equations. April through October data are utilized in these analyses. The results are rejected if they do not meet the following criteria: 0<\( R_{10} \)<15 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) and 0<\( Q_{10} \)<4.

### 3.3.2.1 Seasonal patterns

Seasonal patterns of \( R_{10} \) for each ecosystem are shown in Figure 3.3. As expected, \( R_{10} \) is generally larger in the growing season than in the dormant season. The average values of \( R_{10} \) for the six ecosystem types in the WLEF area vary from 2 to 6 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) in summer. In general, \( Q_{10} \) values are not a strong function of the season, indicating little change in the sensitivity of respiration to temperature across seasons.

### 3.3.2.2 Seasonal average \( R_{10} \)

To compare quantitatively the respiration rates among the ecosystems, the \( Q_{10} \) and \( R_{10} \) in the growing season (May through September) are averaged and presented in Table 3.3. The average value of \( R_{10} \) fitted to the data at the WLEF site is larger than that at the WC site and that at the LC site. \( R_{10} \) observed at the WC site is smaller than those of
the footprint-decomposed mixed coniferous and deciduous forests (type I) and aspen forests (type II) in the WLEF footprint area, and slightly smaller than that for the mature deciduous forest ecosystems (type III). Considering wetland ecosystems, $R_{10}$ at the LC site is smaller than those for both wetland ecosystems (types IV and V) in the WLEF footprint area.

### 3.3.2.3 Seasonal average $Q_{10}$

Seasonally-averaged $Q_{10}$ at the WLEF tower is larger than those at the LC and WC towers. There is no significant difference between the $Q_{10}$ values for the WC and LC sites. Among the upland forest ecosystems (types I, II, and III) in the WLEF footprint area, the aspen ecosystem has the largest $Q_{10}$. The lowland wetland (type IV) has a smaller $Q_{10}$ value than the forested wetland (type V). The forested wetlands have a $Q_{10}$ similar to the upland forest ecosystems. These comparisons imply that the sensitivities of respiration to temperature change may be different among the upland forests with different functional types and species, and between the different wetland types.

In addition, the WC site has a $Q_{10}$ similar to the upland forest ecosystems (types I, II, and III) and the forested wetland ecosystem in the WLEF footprint area. The LC site has a larger $Q_{10}$ than the lowland shrub wetlands (type IV), but the value is similar to the forested wetlands (type V), in the WLEF footprint area, implying that the responses of similar stand types may vary at different sites.
3.3.3 Comparison of daily integrated fluxes

Figure 3.4 shows the May through September average, integrated daily NEE, ecosystem respiration (ER), and gross ecosystem production (GEP) estimated for the six ecosystem types in the WLEF footprint, as well as these same products for the WC, LC and un-decomposed WLEF tower data. The respiration model (Eq. 3.7) and parameters derived using nighttime data are used to calculate daytime ecosystem respiration. NEE values are calculated by using Eqs. (3.5) and (3.6) during the daytime and at night, respectively. The gross ecosystem production is calculated by subtracting the daytime respiration rate from the daytime NEE of CO₂, i.e., \( \text{GEP} = \text{NEE} - \text{ER} \).

3.3.3.1 Average daily ecosystem respiration, ER

Among the three tower sites, the highest and lowest ER values are observed at the WLEF and LC sites, respectively (Figure 3.4a). ER for the upland deciduous forest at the WC site is smaller than ER for the three upland forest stand types (types I, II, and III) in the WLEF footprint area. Aspen shows the highest ER among upland forest types. Forested wetlands (type V) exhibit the highest ER among wetland types. The lowland shrub wetland ecosystem (type IV) has a larger ER than the LC site. The comparison indicates that ER differs among upland forest ecosystem types and between wetland ecosystem types, as well as between similar ecosystem types at different sites.
3.3.3.2 Average daily gross ecosystem production, GEP

The highest GEP (in magnitude) is observed at the WC site, with the lowest value being at the LC site. The WLEF site has an intermediate average daily GEP (Figure 3.4 b) in contrast to its high ER among three tower sites. Aspen (type II) and forested wetland (type IV) exhibit a GEP similar to the WC site. The mixed coniferous and deciduous (type I) and other upland deciduous forest (type III) types have somewhat lower GEP.

Similar to the comparison of ER, the magnitude of the average daily GEP for the lowland shrub wetland ecosystem measured at the LC tower is significantly smaller than that for the same ecosystem type inverted in the WLEF footprint area (type IV), both of which have smaller GEP than the forested wetland ecosystem (type V).

3.3.3.3 Average daily net ecosystem-atmosphere exchange, NEE

Among the three tower sites, the WC site has the highest net uptake of CO$_2$ followed by the LC site. The net uptake of CO$_2$ (|NEE|) measured at WLEF site is the smallest. Among the six ecosystem types in the WLEF footprint area, the mixed coniferous and deciduous ecosystem (type I) has a slightly larger |NEE| than the other two upland forest ecosystems. |NEE| for the forested wetland ecosystem (type V) is larger than that for the lowland shrub wetland ecosystem (type IV) and nearly equal to that for the mixed coniferous and deciduous upland forest ecosystem (types I) in the WLEF footprint area. All forested upland ecosystems and the lowland shrub wetland ecosystem...
(type IV) have smaller |NEE| values in the WLEF footprint area than the ecosystems at the WC site and at the LC site, respectively.

3.3.4 Impacts of uncertainty in surface layer footprint on solutions

The flux footprint model is a critical element of this study. Footprint predictions using the $z_0$ and $d$ values estimated from wind speeds measured at 30 and 122m indicate that surface layer flux measurements are influenced significantly by the grassy area surrounding the tower base, particularly under unstable conditions. But the NEE value measured at the 30m is only slightly smaller than at the two high levels (Ricciuto et al., submitted), which in turn suggests that the footprint predictions are inaccurate assuming that the NEE for the grass around the tower is significantly smaller than that for forests. In addition, the parameters of $z_0$ and $d$ are characteristic of a forest, not grass, contradicting the predictions of the surface layer footprint model using these $z_0$ and $d$ values. The cause of these contradictions is unknown.

To test how sensitive the derived ecosystem model parameters are to the surface layer footprint, we compared those results using different simulated footprints through changing model parameters ($z_0$, $d$, and $\sigma_y$) in the surface layer model. We calculate the footprints for the fluxes measured in the surface layer with the following values of $z_0$ and $d$ in an increasing order of the footprint area at a given stability: (1) $z_0=0.5$m and $d=15$m; (2) $z_0=0.4$m and $d=10$m; (3) $z_0=0.3$m and $d=8$m; (4) $z_0=0.2$m and $d=6$m; (5) $z_0=0.1$m and $d=4$m. Meanwhile, footprints are also computed with two different horizontal diffusion
coefficients in the case of (4), i.e., (6) $z_0=0.2\text{m}$, $d=6\text{m}$, and $\sigma_y$ is half of that in Eq. (3.9); (7) $z_0=0.2\text{m}$, $d=6\text{m}$, and $\sigma_y$ is twice of that in Eq. (3.9).

Figure 3.5 compares the May through September averaged values of $P_{\text{max}}$ and $R$ in the light response model (Eq. 3.5), and $R_{10}$ in the respiration model (Eq. 3.6) inferred using the surface layer footprints in the seven cases. The footprints for measurements above the surface layer remain the same. The differences in the seasonally-averaged $P_{\text{max}}$ among ecosystem types are significant for all cases (Figure 3.5a). Overall, the results are fairly consistent across the range of $z_0$, $d$, and $\sigma_y$ except for case 1. Assuming that the $P_{\text{max}}$ and respiration rate for ecosystem VI (Table 3.1) would be the smallest among all ecosystems, the results of case (4) could be interpretable, which is reported in this paper. It should be pointed out that more advanced models should be used to simulate the footprint more precisely under this complex condition (inhomogeneous surface), which is one of priorities in the future efforts in order to apply the decomposition method described here in practice.

3.4 Discussion

This section examines if fluxes measured at the WLEF tower are representative in the region by testing two hypotheses. Then implications from these tests and from the comparisons among the decomposed fluxes in the WLEF footprint area and measured fluxes at towers are discussed.
3.4.1 Does the WLEF NEE represent the region?

An initial hypothesis is that flux measurements at the WLEF tall tower are representative of the aggregated fluxes in the region since the WLEF footprint covers a large area of mixture of wetland and forested upland stands, apparently encompassing the dominant stand types in the region. This hypothesis can be tested in part by examining the weights of ecosystem types to measured fluxes in the footprint area.

3.4.1.1 Weights of stand types in the footprint area to measured fluxes

As described in Eq. (3.2), eddy-covariance fluxes can be interpreted as the weighted averages of the fluxes from various stand types in the footprint areas. In this study, the weights are a set of integrated footprints over the areas of six identified ecosystem types, as in Eq. (3.4). As an example, Figure 3.6 presents the mean weights for daytime eddy-flux measurements at the three levels of the WLEF tower in the growing season in 2000. The footprints for the measurements at 30m are calculated in the day under near-neutral and unstable conditions, while the footprints for the measurements above the surface layer, i.e., at 122m in some cases and usually at 396m are calculated only under strongly unstable conditions (|L/h|<0.1).

The 30m level is significantly influenced by the grassy clearing at the base of the tower. In contrast, the 122m and 396m footprint areas are mostly covered by the forested upland and wetland ecosystems. The weights of the stand types sampled at 122 and 396m are closer to the fractional areas of the stand types distributed in the region (Table 3.1).
than at 30m, implying that the measurements at the higher levels better represent the area-averaged fluxes in the region. This analysis assumes that the six stand types we have chosen properly represent CO₂ fluxes in the region.

### 3.4.1.2 Possible biases in temporally-integrated NEE

NEE is usually integrated over a time period to examine the responses of the ecosystems to environmental conditions on a variety of temporal scales. In this case, changes in footprint with time can bring difficulties in interpreting the temporally-integrated NEE (e.g., daily or yearly cumulative NEE) over a heterogeneous area. For example, the contribution of the upland grassland near the WLEF tower base to the measured flux is smaller at night than in the day because of flatter and wider footprints under more stable atmospheric conditions (Horst and Weil, 1994). This day and night bias in the influence of the grassy clearing will cause a bias in temporally-integrated NEE (daily or longer), particularly when differences in CO₂ fluxes among stand types are significant. For instance, the daily integrated NEE measured at the WLEF tower is most likely overestimated mathematically (i.e., underestimated net uptake of CO₂) assuming that the grassland has the smaller rates of uptake and release of CO₂ (Note the opposite signs for the two processes) than other stand types but more weight to the flux measurements in the day than at night. This could be one of the explanations why the WLEF tower observes low net uptake of CO₂.

Further, even if the weights of the stand types to measurements in the sampled area were equal to the fractional areas of the respective types (in the region), and the
footprints did not change with time, whether NEE for each stand type in the WLEF footprint area would be representative in the entire region or not is still a critical factor in extending the tower measurements to the region.

3.4.2 Comparison of the ecosystem responses to environmental conditions at the three sites

3.4.2.1 Wetland-upland aggregation hypothesis

The up-scaling hypothesis that the fluxes measured at the WLEF site are area-weighted averages of the fluxes measured at the LC and WC sites is utilized to assess whether the measurements at the two short towers can represent the NEEs for all the wetlands and uplands in the region. In this hypothesis, all ecosystems are categorized into upland or wetland ecosystems based on watershed functions (Table 3.1).

According to this hypothesis, the average daily integrated GEP and ER at the WLEF site should be an intermediate value between those measured at the LC and WC sites, respectively. However, the data show that the average daily ER measured at the WLEF site is larger than either of the other two sites (Figure 3.4). Consequently, there is no way the weighted mean of the observed ER at the WC and LC towers can equal that at the WLEF tower. The hypothesis must be rejected unless the error in the measurements is greater than the difference (unlikely in this case). The same conclusion is drawn from the respiration rates estimated by fitting the two ecosystem models (i.e., $R_d$ in Eq. 3.5 and
$R_{fo}$ in Eq. 3.6) to the data measured at the WLEF tower and the other two towers (Tables 3.2 and 3.3). The hypothesis cannot be rejected, however, for the average daily-integrated GEP because the GEP at the WLEF is in between those at the two other sites.

The rejection of this hypothesis indicates that the watershed-function level of classification is insufficient for upscaling in the region. In other words, the response of some stand types within the category of wetland, or within the category of forest upland, or both, to the same environmental conditions can be significantly different from one another, particularly in terms of respiration rates. Another implication from the rejection is that the fluxes could be different for the same stand types at different sites. Both implications are supported partially by the comparisons of the fluxes and functional parameters among stand types and among sites in section 3.3.

### 3.4.2.2 Six stand type aggregation hypothesis

Since the watershed function classification scheme is obviously insufficient, we hypothesize that regional fluxes can be described as an aggregate of fluxes for six different stand types, which is motivated in part by Mackay et al. (Mackay et al., 2002). We evaluate this hypothesis by comparing the WLEF flux from the footprint decomposition to the WC and LC flux measurements.

The ER values for stand types around the WLEF tower are larger than the stands of the same type sampled by the WC and LC towers. GEP values for the upland deciduous forests in the WLEF footprint area are smaller than that at the WC site. Therefore, the magnitude of NEE at the WC site is larger than those of the upland
deciduous forests around WLEF by a factor of about 3. These comparisons suggest that the measurements at the WC site (a mature upland deciduous forest stand) are not representative of those of the upland forests in the WLEF footprint area. Unlike ER, GEP measured at the LC site is similar to that for the lowland wetlands in the WLEF footprint area. As a result, NEE at the LC site is more negative and cannot represent the lowland wetland ecosystem in the WLEF footprint area. These results suggest that a more detailed classification scheme is needed for both wetland and upland deciduous forest stand types. Another possible explanation is due to errors in the flux decomposition, but the ER values are so different that it seems unlikely that such errors would resolve the discrepancies. Therefore, the upscaling scheme based on the six stand type classification is still improper.

The high ER measured at the WLEF tower might be in part due to the effects of the surrounding forested wetland and aspen ecosystems whose respiration rates are higher than others. The understory is sparse at the WC site compared to the WLEF site, likely accounting in part for the significant differences in the ER for the ecosystems with the upland deciduous forests at the two sites. Other reasons we do not rule out include the effects of other factors including canopy age and density, tree roots, soil, and litter quality on the ecosystem respiration. Differences in GEP might be due to plant species, canopy density, micrometeorological conditions, and other biological factors. More observations are needed to specially identify the differences in stands that are critical to flux aggregation. Determining how to properly classify the ecosystem and design measurements is a challenge for the bottom-up method.
3.5 Aggregation experiments

Despite our inability to present a landscape classification scheme that reconciles WLEF flux data with those from WC and LC, it is instructive to construct a variety of aggregated flux measurements using these data. These aggregations present a quantitative measure of the uncertainty in regional fluxes that arises from the insufficient classification schemes. Spatially aggregated estimates of ER, GEP, and NEE are developed from the regional vegetation map and four different combinations of the ecosystem type fluxes and tower measurement. The four aggregation approaches are:

1. Watershed-function level classification. The ecosystems of the regional vegetation map are classified into two types, wetland or upland. Wetland fluxes are represented by LC site measurements and upland fluxes are represented by WC site measurements.

2. Stand type level classification. The inferred fluxes for the six ecosystem types in Table 3.1 in the WLEF footprint area are used to represent the entire region.

3. The same scenario as case (2) with the exception that WC data are used in place of the inferred flux for the upland deciduous forest ecosystem type flux (type III) and LC data are used in place of the inferred flux for lowland shrub wetlands (type IV).

4. Integrated WLEF fluxes. The fluxes measured at the WLEF tower are integrated directly and assumed to be regional estimates.

Figure 3.7 compares the averaged daily-integrated ER, NEE, and GEP for each method during the growing season in 2003. The aggregated daily ER is the lowest in Method 1, while no significant difference is found among the other three methods. The
magnitude of GEP estimated directly by integrating the WLEF tower data with time (Method 4) is the smallest. There is no significant difference in GEP among the three aggregations. The large ER but small GEP result in the least negative NEE (smallest net uptake) from direct integration of the WLEF tower flux data (Method 4). The large GEP and small ER result in the most negative NEE (largest net uptake) in Method 1. We hypothesize that the results from aggregation method 3 are the best estimates for the regional fluxes as the broadest sample of observations are included. Unless other stand types covering large areas and having fluxes falling out of the range of those presented here exist, aggregations 1 and 4 suggest bounds for the regional fluxes; the difference in the cumulative regional NEE estimate between these two aggregations is large, about 400gC m$^{-2}$ over the growing season. This large difference found among the small number of flux observations suggests that the range of fluxes among stands in this region is greatly under-sampled in this study. This also shows the danger of using any single flux tower to describe a large region.

3.6 Limitations and future work

The decomposition method developed here has several limitations. First, the accuracy of the classification inherent in the vegetation map is potentially important and has not been taken into account in this analysis. The overall accuracy is in the range of about 40% to 90% depending on classes (WiDNR, 1998). Misclassification of stand type will cause errors in the flux decomposition. The land cover data product was derived from LANDSAT Thematic Mapper (TM) satellite imagery acquired from fly-overs over
a period of years (WiDNR, 1998). As a result, the data cannot reflect recent changes in vegetation caused by ongoing forest management and manipulation of water table depth. A more accurate inventory of the stand types would improve the representativeness of any aggregation and the interpretation of the measurements at the WLEF tower.

Second, satellite remote sensors have difficulty detecting forest understory, litter, vegetation density, stand age, and stand height, all of which may be needed to describe the carbon exchange between forests and the atmosphere. It appears that ecosystem fluxes are dependent not only on stand type but also on those additional stand features, as suggested by the comparison between the fluxes of the same stand types at different locations in this study and as shown with additional stand-level flux measurements by Desai et al. (submitted). Consequently, to aggregate the measured fluxes at towers more effectively, we need to consider classification schemes, which go beyond land cover focused only on plant functional types and take into account additional ecosystem variables. This study does not have enough data to define which variables must be included. Identifying the variables needed and mapping them accurately are challenges to the aggregation of carbon flux measurements.

Third, the precision and accuracy of the derived functional parameters and fluxes are limited by the uncertainty of the simulated footprints, the ecosystem models, and flux measurements. The underlying flux decomposition is sensitive to the uncertainty of the eddy flux measurements, an uncertainty that is also difficult to evaluate accurately despite many discussions in the literature. Errors in the derived parameters can be resulted from the use of the equal weights for all measured fluxes in the calculation. The footprint models apply only over dynamically-homogeneous surfaces; this is usually not
met in practice. The footprint uncertainty introduced by applying the models over inhomogeneous surfaces cannot be quantified. In addition, the effects of change in wind direction with height on footprint estimates are not taken into account.

Nevertheless, the ability to derive fluxes and ecosystem parameters that show reasonable distinctions among the stand types suggests that this decomposition method has succeeded in this application despite the numerous sources of uncertainty. Addressing the sources of uncertainty would increase confidence in these results. More precisely, advanced footprint models considering heterogeneous flow and surface conditions should be used to simulate the footprints in the future. With respect to the photosynthesis and respiration models, the effects of soil moisture are needed to take into account to describe more precisely the responses of the ecosystem types to environmental conditions. Soil moisture is the most difficult to consider in that it is recorded only at a single point; the measurement is unlikely to represent both upland and wetland ecosystems well. Further studies will require more measurements, e.g., spatial distribution of soil moisture, to be incorporated into the model.

Using the decomposition method shown here is impractical if the number of the stand types that must be distinguished is large. Larger number of stand type result in more unknowns, and the requirements of accuracy for the footprints, vegetation map, and flux measurements becomes higher and more difficult to meet. In practice, this method is likely to be used in combination with other measurements or modeling (e.g., Desai et al., submitted). A complementary and independent approach that somewhat avoids these problems is a regional atmospheric budget which can provide additional constraints for the aggregated fluxes (see chapter 4).
3.7 Summary and conclusions

An approach has been developed with the aid of footprint and ecosystem models to estimate the flux of each stand type in a footprint area by decomposing tower-based eddy-covariance fluxes. CO₂ fluxes for six stand types are inferred from the measurements at the WLEF tower in northern Wisconsin. Ecological parameters and fluxes are derived for each of the six stand types. The fluxes and functional parameters of the six stand types in the WLEF footprint area are compared with each other and with those observed at the WC and LC sites. Regional fluxes with different aggregation schemes are compared. Conclusions are as follows.

1. The difference in NEE measured at the WC and LC sites is about 2.8 gC m⁻² day⁻¹ during the growing season, indicating that it is critical to distinguish wetlands from uplands for flux aggregation. Moreover, within the broad category of upland deciduous forests, aspen stands respond differently than other deciduous hardwoods to the same environmental conditions. There is also a distinction between lowland shrub wetlands and forested wetlands. Aspen stands and forested wetlands have high GEP and ER. Confirmation of these differences by direct measurements would support the findings here. CO₂ fluxes measured over the deciduous forest at WC and the wetland at LC are compared with those with the same stand types in the WLEF footprint area. The comparison shows higher GEP and lower ER at the WC site, and slightly higher GEP and ER at the LC site as compared to similar stands around WLEF. The differences imply that the six stand classification scheme does not capture all the variability in stand
characteristics relevant to CO$_2$ exchange. With and without the stand type 
classification scheme and the difference between sites being considered, the 
difference in the (cumulative) aggregated NEE can be as large as 250gCm$^{-2}$ over 
the growing season. The large respiration rates observed at the WLEF tower 
might be due in part to the contribution of the aspen and forested wetland 
etcosystems in the footprint area. In addition, the difference in the respiration rate 
for the same stand types at different sites implies the flaw of the currently-used 
etystem classification scheme.

(2) The hypothesis that the respiration rate measured at the WLEF tower could be 
explained by an area-weighted average of respiration rates measured at the WC 
and LC towers is rejected, proving that the watershed-function level of the 
etystem classification is insufficient for flux aggregation in this region. 
Footprint modeling suggests that flux measurements at the WLEF tower are not 
representative of the region in terms of the weight of each stand type to the 
measured flux in the footprint area in comparison with the fractional coverage of 
the corresponding stand type in the region. It is hard to interpret the temporally-
integrated NEE measured at the WLEF tower without the footprints being 
quantified because the fractional contribution of each stand type in the footprint 
area to the measurements changes with time.

(3) Regional flux estimates from four approaches are different, indicating a 
quantitative measure of the uncertainty in regional fluxes that arises from the 
insufficient classification schemes. In the growing season, the aggregated daily 
NEE values are about -1.79±0.51gC m$^{-2}$ (error bar is one standard error) using
the stand-type classification scheme, $-3.51 \pm 0.27 \text{gC m}^{-2}$ using the watershed-function classification scheme (wetland and upland), and $-0.38 \pm 0.13 \text{gC m}^{-2}$ directly integrated from the WLEF data. The maximum difference in the (cumulative) regional NEE estimate can be as large as about $400 \text{gC m}^{-2}$ over the growing season.

The ability to derive fluxes and ecosystem parameters that show reasonable distinctions among the stand types suggests that this decomposition method has succeeded in this application despite the numerous sources of uncertainty. The precision and accuracy of the results are limited by the accuracy of the vegetation map, the simulated footprints, ecosystem models, and flux measurements. Quantifying those errors, especially the vegetation map and footprints, is one of important goals of future work. Flux footprint research should be extended to heterogeneous terrain and ecosystems. More measurements are needed to interpret NEE measured over the heterogeneous ecosystem.
Figure 3.1: Distribution of land cover classes in the 60km×60km region centered on the WLEF tower. The three pluses represent the locations of the WC, LC and WLEF towers. Data source: WiscLAND (WiDNR, 1998). The flux aggregation is conducted in the 40km×40km area centered at WLEF.
Figure 3.2: Monthly variation of the inferred functional parameters in the light-response model (Eq. 3.5) for the six ecosystem types in the WLEF footprint area in the growing season in 2000. The parameters fitted to the WC and un-decomposed WLEF tower data (dotted lines) are also shown for comparison. The error bars are the standard deviations of the means.
Figure 3.3: Monthly variation of the inferred functional parameters in the respiration model (Eq. 3.6) for each ecosystem type in the WLEF footprint area from April to October in 2000. The parameters fitted to WC and un-decomposed WLEF data are also shown for comparison. This analysis uses only nighttime flux data. The error bars are the standard deviations of the means.
Figure 3.4: The average daily-integrated ecosystem respiration flux (a), gross ecosystem production (b), and NEE (c), over the period of the growing seasons (May through Sep) in 2000 and 2003. Note that the Lost Creek data are not available in 2000. The error bars are the standard deviations of the means.
Figure 3.5: May through September averaged parameters of (a) $P_{\text{max}}$, (b) $R_d$ in the light response model, and (c) $R_{10}$ in the respiration model vary when the surface layer footprints are estimated using different $z_0$ and $d$ values (see text). Case 1: $z_0=0.5\text{m}$, $d=15\text{m}$; case 2: $z_0=0.4\text{m}$, $d=10\text{m}$; case 3: $z_0=0.3\text{m}$, $d=8\text{m}$; case 4: $z_0=0.2\text{m}$, $d=6\text{m}$; case 5: $z_0=0.1\text{m}$, $d=4\text{m}$; case 6: $z_0=0.2\text{m}$, $d=6\text{m}$, $\sigma_y$ is reduced by a factor of 2; case 7: $z_0=0.2\text{m}$, $d=6\text{m}$, $\sigma_y$ is doubled. Note that footprints for the measurements above the surface layer remain the same in all cases.
Figure 3.6: The mean weights of each stand type for daytime flux measurements from the three levels of the WLEF tower during May through September of the year 2000. The 30 m footprints are calculated under near-neutral to unstable conditions. When the acceptor heights are above the surface layer, as is often the case for 122 m and 396 m flux measurements, footprints are only computed for strongly unstable conditions. Each error bar represents the standard deviation of the calculated hourly weights during the season.
Figure 3.7: Daily-integrated, aggregated fluxes averaged over the growing seasons (May through September) in 2003. (a) ER, (b) GEP, and (c) NEE for different approaches (see section 3.5). The error bars are the standard deviations of the means.
Table 3.1: Ecosystem classification

<table>
<thead>
<tr>
<th>Level 1-watershed function level</th>
<th>Level 2 – stand type level</th>
<th>Fractional area&lt;sup&gt;6&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>Watershed functions</td>
<td>Ecosystem Types</td>
<td>Land cover classes included</td>
</tr>
<tr>
<td>Forested upland</td>
<td>I Coniferous, mixed coniferous and deciduous, e.g., jack pine, red pine and white spruce, etc.</td>
<td>14.3%</td>
</tr>
<tr>
<td></td>
<td>II Aspen</td>
<td>18.5%</td>
</tr>
<tr>
<td></td>
<td>III Other upland deciduous e.g., Oak, Maple, etc.</td>
<td>17.4%</td>
</tr>
<tr>
<td>Wetland</td>
<td>IV Lowland shrub wetland&lt;sup&gt;6&lt;/sup&gt;</td>
<td>18.0%</td>
</tr>
<tr>
<td></td>
<td>V Forested wetland&lt;sup&gt;55&lt;/sup&gt;</td>
<td>16.2%</td>
</tr>
<tr>
<td>Others</td>
<td>VI Other land cover classes: urban/developed, agriculture, grassland, open water, shrub land, and barren</td>
<td>15.6%</td>
</tr>
</tbody>
</table>

<sup>5</sup>Lowland Shrub wetland: Woody vegetation, less than 20 feet tall, with a tree cover of less than 10%, and occurring in wetland areas.<sup>55</sup>Forested Wetland: Wetlands dominated by woody perennial plants, with a canopy cover greater than 10% , and trees reaching a mature height of at least 6 feet (WiDNR, 1998). <sup>6</sup>Fractional areas of the six ecosystems in the 40×40km<sup>2</sup> region centered at the WLEF tower.
Table 3.2: The June to August 2000 and 2003 averages and standard errors of the three parameters, $\alpha$, $P_{\text{max}}$, and $R_d$, in the light-response model (Eq. 3.5) for the six ecosystem types in the WLEF footprint area. The parameters fitted to the WLEF, WC and LC data (with no footprint decomposition for the WLEF results) are also shown. CO$_2$ fluxes for open water and roads are assumed to be zero.

<table>
<thead>
<tr>
<th>Stand type</th>
<th>$\alpha$ $\mu$molC/$\mu$mol quanta</th>
<th>$P_{\text{max}}$ $\mu$molC/m$^2$/s</th>
<th>$R_d$ $\mu$molC/m$^2$/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Mixed Coniferous/Deciduous</td>
<td>-0.051±0.004</td>
<td>-28.6±1.8</td>
<td>6.4±0.3</td>
</tr>
<tr>
<td>II. Aspen</td>
<td>-0.046±0.004</td>
<td>-38.9±2.4</td>
<td>7.1±0.3</td>
</tr>
<tr>
<td>III. Other Deciduous</td>
<td>-0.054±0.004</td>
<td>-24.5±1.4</td>
<td>6.8±0.4</td>
</tr>
<tr>
<td>IV. Lowland Wetland</td>
<td>-0.052±0.004</td>
<td>-23.5±1.9</td>
<td>6.4±0.3</td>
</tr>
<tr>
<td>V. Forested wetland</td>
<td>-0.048±0.003</td>
<td>-34.8±2.3</td>
<td>6.2±0.3</td>
</tr>
<tr>
<td>VI. Others</td>
<td>-0.056±0.003</td>
<td>-13.5±0.6</td>
<td>5.0±0.2</td>
</tr>
<tr>
<td><strong>WLEF tower</strong></td>
<td><strong>-0.051±0.004</strong></td>
<td><strong>-22.9±1.2</strong></td>
<td><strong>6.0±0.4</strong></td>
</tr>
<tr>
<td><strong>Willow Creek tower</strong></td>
<td><strong>-0.055±0.002</strong></td>
<td><strong>-37.3±1.1</strong></td>
<td><strong>4.9±0.2</strong></td>
</tr>
<tr>
<td><strong>Lost Creek tower</strong></td>
<td><strong>-0.055±0.003</strong></td>
<td><strong>-12.5±0.2</strong></td>
<td><strong>3.5±0.1</strong></td>
</tr>
</tbody>
</table>
Table 3.3: The averages and standard errors of the two functional parameters, $R_{10}$ and $Q_{10}$, in the respiration model (Eq. 3.6) for each ecosystem type in the growing seasons (May through September) in 2000 and 2003. The parameters fitted to the two stand-level towers (WC and LC) and the un-decomposed WLEF data are also shown. Only nighttime flux data are used in this analysis. CO$_2$ fluxes for open water and roads are assumed to be zero.

<table>
<thead>
<tr>
<th>Ecosystem type or tower site name</th>
<th>$R_{10}$ $\mu$molC/m$^2$/s</th>
<th>$Q_{10}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Mixed Coniferous/Deciduous</td>
<td>3.2±0.25</td>
<td>1.7±0.10</td>
</tr>
<tr>
<td>II. Aspen</td>
<td>3.7±0.25</td>
<td>1.9±0.10</td>
</tr>
<tr>
<td>III. Other Deciduous</td>
<td>3.1±0.34</td>
<td>1.7±0.09</td>
</tr>
<tr>
<td>IV. Lowland Wetland</td>
<td>2.9±0.20</td>
<td>1.6±0.05</td>
</tr>
<tr>
<td>V. Forested wetland</td>
<td>3.6±0.40</td>
<td>1.9±0.11</td>
</tr>
<tr>
<td>VI. Others</td>
<td>2.7±0.23</td>
<td>1.5±0.07</td>
</tr>
<tr>
<td>WLEF tower</td>
<td>3.4±0.14</td>
<td>2.0±0.08</td>
</tr>
<tr>
<td>Willow Creek tower</td>
<td>2.7±0.06</td>
<td>1.8±0.08</td>
</tr>
<tr>
<td>Lost Creek tower</td>
<td>2.0±0.15</td>
<td>1.9±0.14</td>
</tr>
</tbody>
</table>
Chapter 4

Inferring Long-Term Regional-Scale CO₂ Fluxes over a Forested Area in Northern Wisconsin using CO₂ Mixing Ratio Measurements from a Very Tall Tower

4.1 Introduction

Terrestrial ecosystems play a critical role in buffering the climate change caused by the carbon dioxide (CO₂) emitted from fossil fuels burning (IPCC, 2001). Quantifying the net ecosystem-atmosphere exchange of CO₂ (NEE) over various temporal and spatial scales improves our understanding of the terrestrial carbon budget. Estimates of NEE are well-constrained at the global scale by means of inverse modeling (Bousquet et al., 1999a; Bousquet et al., 1999b; Enting, 2002; Gurney et al., 2002; Rayner et al., 1996), and NEE can be directly observed at a local scale via the eddy-covariance method (Baldocchi et al., 2001). It is more difficult to infer NEE at intermediate scales, and this has limited our understanding of the interaction between climate change and terrestrial ecosystems. As a result, significant uncertainty exists in predictions of future changes in climate and atmospheric CO₂ concentrations (Cao and Woodward, 1998; Cramer et al., 2001; Cramer et al., 1999; Huntingford et al., 2000; IPCC, 2001; Schlesinger, 1983). No standard approaches have yet been developed to estimate NEE on intermediate scales.
Tower-based eddy covariance systems can measure NEE on long time scales and spatial scales of the order of a few km$^2$ (Baldocchi et al., 2001). These measurements may not be representative of a larger region particularly when the region is heterogeneous. NEE over larger areas can be estimated by aircraft measurement techniques (e.g., Barr et al., 1997; Oncley et al., 1997), but this method is expensive and restricted to short time scales such as hours to days.

Flux aggregation is another approach used to up-scale measurements from local to regional scales. A major challenge of this approach is whether measurements at one site with an ecosystem type are representative of the region. This issue arises not only due to the quality of the measurements themselves, but also due to the ecosystem classification scheme used. Evidence in chapter 3 and other studies (e.g., Desai et al., submitted) suggests that the usually-used land-cover-based classification scheme may be insufficient for aggregating ecosystem fluxes in northern Wisconsin. This is probably because, excluding climate, NEE variability is governed not only by vegetation type but also by stand age and density, soil type, and many other factors (Binkley et al., 2002; Chen et al., 2002; Desai et al., submitted; Litton et al., 2004; Litvak et al., 2003; Ryan et al., 2004) that cannot be described by the land-cover-based classification scheme. It is impractical to make observations for each ecosystem type if the number of the ecosystem types is large. The impacts of nonlinear processes such as edge effects on NEE estimates are also difficult to be accounted for with the aggregation approach.

The atmospheric boundary layer (ABL) budget technique is a complementary approach and based on mass conservation in the ABL. In the daytime convective boundary layer (CBL), this method takes advantage of CBL bulk properties that are
independent of small scale (e.g., <10-20km) surface heterogeneities. Advection by the mean flow as well as mixing by large eddies in the CBL naturally aggregate fluxes of CO\textsubscript{2} that are emitted from or absorbed by small scale processes across different surfaces in a region (Raupach et al., 1992). In this case, the average surface flux over a region can be estimated provided that all components in the budget equation are measurable, predictable or negligible. Raupach et al. (1992) showed that the length scale of such a regional aggregation is typically 5-30km, which is larger than footprint scales of fluxes measured via eddy-covariance at short towers. Recently a number of case studies on the applications of the CBL budget method to estimating surface fluxes have been reported (Gryning and Batchvarova, 1999; Helmsg et al., 1998; Kuck et al., 2000; Levy et al., 1999; Lloyd et al., 2001; McNaughton and Spriggs, 1986; Raupach et al., 1992), but the time scales of those applications are short because the field campaigns usually last for a short period such as a few days. Some quantities necessary for budget calculations are not routinely observed, including the CO\textsubscript{2} mixing ratio in the mixed layer and just above the CBL, the horizontal gradient of the CO\textsubscript{2} mixing ratio within the CBL, the CBL depth, and the large scale vertical velocity at the top of CBL. Therefore the components in the budget equation for CO\textsubscript{2} are not easy to construct in practice except in special short-term campaigns. For example, the budget method has been applied to estimate CO\textsubscript{2} surface fluxes using an aircraft (Lloyd et al., 2001) or tethered balloon (Kuck et al., 2000) measurements.

Multi-level measurements of CO\textsubscript{2} mixing ratios at tall towers that are higher than the surface layer can make the budget method easier to apply in practice because the vertical profile of CO\textsubscript{2} mixing ratio can be directly and continuously measured from the
surface to the mixed layer for long periods of time. In this case, fewer assumptions are required than the cases when measurements are made only near the surface. For example, we do not need to use the similarity theory to infer CO2 mixing ratio in the mixed layer. As an application of tall tower measurements, we take advantage of the measurements of CO2 mixing ratios at six levels of a 447-m tall tower (WLEF) in this study. More importantly, these continuous measurements of the CO2 mixing ratio profile from the surface layer to mid-CBL can provide regional NEE estimates on a longer time scale compared with those reported in the literature.

It is more difficult to use the budget method to infer surface fluxes in the nocturnal boundary layer (NBL) than in the daytime CBL. In the literature, only a few such applications are reported (Denmead et al., 1996; Pattey et al., 2002). The turbulence in the NBL is sometimes patchy and sporadic under very stable conditions (Stull, 1988), resulting in poor representativeness of the measurements from a single location. Therefore, fluxes inferred from the budget method with the measurements at the single location can be possibly representative of a larger area only when the NBL is well-mixed. In addition, the depth of the NBL is not easy to quantify compared with the CBL depth, hence restricting the application of the NBL budget method. In this study, we attempt to use the vertical profiles of CO2 mixing ratios at the 447-m tall tower to avoid predicting NBL depth directly, and an approach is proposed to identify and remove the cases when systematic errors or non-representative sampling occurs.

The structure of this chapter is as follows. The site and methods under study are described in section 4.2. Results are presented in section 4.3, along with a discussion of uncertainty. Results include the inferred daytime and nighttime NEE values during the
growing and dormant seasons, their dependences upon environmental variables such as radiation and temperature, and comparison with measurements from three eddy-covariance towers. Finally, a summary is offered in section 4.4.

4.2 Materials and methods

4.2.1 Site and Data

The study region is a 40×40km² region, centered at a tall communication tower (WLEF) (45.9455878° N, 90.272304°W), and is located in the Chequamegon National Forest. The tower is located about 15 km east of Park Falls, Wisconsin. The region is relatively flat with patches of wetland and uplands with mixed evergreen and deciduous forests.

Fluxes of momentum, sensible heat, latent heat, and CO₂ are measured with the eddy-covariance method at three towers in the region: Willow Creek (WC), Lost Creek (LC), and WLEF towers (see Figure 3.1). The WLEF tower is part of the National Oceanographic and Atmospheric Administration's tall tower CO₂ monitoring network (Bakwin et al., 1998). Detailed descriptions of the tower sites and measurements can be found in the literature (Berger et al., 2001; Cook et al., 2004). At WLEF, fluxes are measured at three levels (30, 122, and 396m above the ground level), and CO₂ mixing ratio data, traceable to WMO primary standards, are collected at six levels (11, 30, 76, 122, 244, and 396m, above the ground level) (Bakwin et al., 1998; Davis et al., 2003).
These high-altitude profiles of CO$_2$ flux and mixing ratio are unique among the current AmeriFlux sites. Measurements began in 1995. The land cover types in the typical daytime footprint areas for eddy-covariance CO$_2$ flux measurements at the top two levels of the WLEF tower are mainly mixed wetland and upland forests. The grassland around the tower base contributes significantly to flux measurements at the bottom level under unstable atmospheric conditions according to footprint modeling studies (see chapters 2 and 3). The WC tower is located about 20km southeast of the WLEF tower. The footprint area of flux measurements at this tower is mainly covered by upland deciduous forests. The LC tower is located in a wetland area, about 25km northeast of the WLEF tower. Flux measurements at the WC and LC towers began in 1999 and 2001, respectively.

Other datasets in addition to tower-based data are used to construct the ABL budget components or test simplifying assumptions. A 915MHz boundary-layer profiler radar was deployed in the region for two years (1998 and 1999) to study the daytime CBL. Yi et al. (2001) used these measurements to develop a semi-empirical model to predict the CBL depth using flux measurements. The vertical velocity at the top of the daytime convective boundary layer is estimated from the Rapid Updated Cycle (RUC) reanalysis data (Benjamin et al., 2004) because direct measurements are unavailable. The vertical distribution of CO$_2$ mixing ratio from the upper boundary layer to lower troposphere is obtained from aircraft measurements supported by the NOAA CMDL Carbon Cycle Aircraft Sampling Program (Tans, 1996), which is used to estimate CO$_2$ mixing ratio above the CBL in this study. The air samples are generally collected biweekly at different altitudes during a single flight using an automated Programmable Flask Package. In the case of missing aircraft data (mostly in 2000), free-troposphere
CO₂ mixing ratios are approximated as the average of the CO₂ mixing ratios measured in the marine boundary layer at a downstream site (AZR, 38.77°N, 27.38°W) and in the free troposphere at an upstream inland aircraft profiling site (CAR, 40.9°N, 108.8°W). CO₂ mixing ratio profiles through the boundary layer and into the free troposphere were also measured at this site using a powered parachute platform (Schulz et al., 2004). The flights were conducted hourly from morning to afternoon under fair weather conditions in May and October, 2001, and August, 2002, providing data about the evolution of CO₂ mixing ratio with time both within and above the CBL.

4.2.2 Equations and assumptions

This section describes the equations and assumptions used in the calculation of the components of the CBL and NBL CO₂ budgets.

4.2.2.1 In the CBL

Averaging the CO₂ mixing ratio conservation equation over a horizontal area yields,

\[
\frac{\partial [c]}{\partial t} + [u] \frac{\partial [c]}{\partial x} + [v] \frac{\partial [c]}{\partial y} + [w] \frac{\partial [c]}{\partial z} + \frac{\partial [uc]}{\partial x} + \frac{\partial [vc]}{\partial y} + \frac{\partial [wc]}{\partial z} = [S],
\]

(4.1)

where \([u]\), \([v]\), and \([w]\) denote the area-averaged wind velocities in \(x\), \(y\), and \(z\)-direction, respectively; \([c]\) and \([S]\) are the area-averaged mixing ratio of CO₂ and (canopy) source strength, respectively; \([uc]\), \([vc]\), and \([wc]\) are the sub-grid scale (smaller than the area
over which the equation is averaged) fluxes in the x-, y-, and z-direction, respectively. Under fair weather conditions, the CBL is divided vertically into three layers as schematically shown in Figure 4.1, namely, the surface layer (0-\(h_0\)), mixed layer (from \(h_0\) to \(h_1\)), and interfacial layer (from \(h_1\) to \(h_2\)). Integrating Eq. (4.1) from the surface to the top of CBL, and ignoring the horizontal sub-grid flux divergence terms (Denmead et al., 1996; Kuck et al., 2000; Levy et al., 1999), we have,

\[
\int_0^{h_2} \frac{\partial [c]}{\partial t} dz + \int_0^{h_2} \left( \left[ [u] \frac{\partial [c]}{\partial x} + [v] \frac{\partial [c]}{\partial y} \right] dz + \int_0^{h_2} \frac{\partial [c]}{\partial z} dz + \int_0^{h_2} \frac{\partial [wc]}{\partial z} dz \right) dz = \int_0^{h_2} [S]dz. \tag{4.2}
\]

Using the Leibnitz rule of integration and rearranging Eq. (4.2) yield,

\[
[NEE] \equiv \left[ wc \right]_0 + \int_0^{h_2} [S]dz,
\]

\[
= h_2 \frac{\partial [c]_{m}}{\partial t} + \left( [c]_{m} - [c]_{h_2} \right) \frac{\partial h_2}{\partial t} + \text{hor}_-\text{adv} + \text{ver}_-\text{adv}
\]

where \([NEE]\) is the area-averaged NEE; \(z_c\) is the canopy height and we assume that no source or sink of CO\(_2\) exists above the canopy; \(\text{hor}_-\text{adv}\) and \(\text{ver}_-\text{adv}\) are the horizontal and vertical advection terms, respectively, i.e. the second and third integrals on the left-hand side of Eq. (4.2). \([c]_{h_2}\) is the mixing ratio of CO\(_2\) just above the interfacial layer \(h_2\); \([c]_{m}\) is the mean mixing ratio of CO\(_2\) in the CBL, which can be expressed in terms of the mean mixing ratios respectively in the three layers, i.e.,

\[
[c]_m = \frac{1}{h_2} \int_0^{h_2} [c]dz = \frac{1}{h_2} \left[ \int_0^{h_2} [c]dz + \int_0^{h_1} [c]dz + \int_{h_1}^{h_2} [c]dz \right], \tag{4.4}
\]

\[
= \frac{1}{h_2} \left( h_0[c]_{m_0} + (h_1 - h_0)[c]_{m_1} + (h_2 - h_1)[c]_{m_2} \right)
\]
where \([c_{m0}], [c_{m1}], \text{and} [c_{m2}]\) are the mean mixing ratios in the surface layer, mixed layer, and interfacial layer, respectively. In deriving Eq. (4.3), the vertical sub-grid scale flux, \([wc]\), at the top of interfacial layer has been assumed to be negligible. \([wc]\) can be interpreted as the sum of the turbulent flux and the mesoscale (the scale between turbulent scale (1km) and the averaged area scale) flux. The former is weak above the CBL top compared with the surface flux. The latter can also be ignored after being averaged over a long time because meso-scale fluxes are usually random (personal communication, Larry Mahrt et al.) unless they are induced by stationary circulations on the grid scale due to, e.g., significant contrasts of surface characteristics like land and water. The landscape around the WLEF tower is not likely to have any large, strong contrasts in the surface energy balance that could drive persistent mesoscale circulation during the day. With the mean-value theorem, the vertical advection term can be rewritten as,

\[
\int_{h_0}^{h_1} [w] \frac{\partial [c]}{\partial z} dz = [w]_{h_0} \{[c]_{h_2} - [c]_{h_1}\} + [w]_{h_1} \{[c]_{h_0} - [c]_{h_0}\}, \tag{4.5}
\]

where \(h_1\) and \(h_0\) are two heights between \(h_1\) and \(h_2\) and between surface and \(h_0\), respectively; \([c]_0\), \([c]_{h0}\), and \([c]_{h1}\) are the averaged mixing ratios of CO2 on the surface, the surface layer top, and the mixed layer top, respectively. The effect of the vertical advection is ignored in the mixed layer due to the small gradient of the CO2 mixing ratio compared with that in the interfacial layer. The second term on the right-hand side (RHS) of the above equation is also ignored because the regional average vertical velocity near the surface is usually small, though it is hard to measure.
With the assumptions of horizontal homogeneity, zero depth of the interfacial layer (i.e., \( h_1=h_2=h \)), and negligible surface layer depth \( (h_0=0) \), Eq. (4.3) reduces to,

\[
[NEE] = h \left( \frac{\partial [c]_m}{\partial t} - \left( [c]_m - [c] \right) \left( \frac{\partial h}{\partial t} - w_+ \right) \right), \tag{4.6}
\]

where \( h \) is the CBL depth; \( w_+ \) is the vertical velocity at the CBL top; \([c]_m\) is the mixing ratio just above the CBL top. Eq. (4.6) is the one-dimensional, zero-order jump CBL model often used in the literature to infer CO\(_2\) fluxes (Denmead et al., 1996; Kuck et al., 2000; Levy et al., 1999; Lloyd et al., 2001; Raupach et al., 1992).

### 4.2.2.2 In the NBL

Under the assumption that the advection terms are negligible after long term averaging in the budget equation, nighttime NEE, representing only respiratory processes in the ecosystem, can be estimated from (Denmead et al., 1996; Pattey et al., 2002),

\[
[NEE] = \int_0^{h_s} \left\{ \frac{\partial [c]}{\partial t} \right\} dz, \tag{4.7}
\]

where \( h_s \) is the height of the top of the NBL below which the CO\(_2\) released by ecosystems is contained; The plus represents just above the NBL top. Due to difficulties in quantifying \( h_s \) in practice, it is replaced by the highest measurement level of CO\(_2\) mixing ratio at the tall tower, i.e., 396m, in the following calculation because NBL depth is usually tens or hundreds of meters and smaller than 396m in many cases. This budget estimates is invalid when the NBL depth is greater than 396m. Criteria that define when Eq. (4.7) is valid will be defined later in terms of atmospheric stability.
4.3 Results and discussions

This section presents typical values of the terms in the budget equation and the seasonally-averaged NEE derived from the budget method using the mixing ratio measurements at the tall tower. The estimated NEE and derived ecosystem parameters are compared with those estimated using the eddy-covariance system at three levels of the tall tower and at WC and LC towers. We define the growing and dormant seasons as the periods from May to September and from October to April, respectively. Data from 2000 and 2003 are used. These data were chosen because fluxes of CO$_2$ and water vapor were affected by an outbreak of tent caterpillar in 2001 and a data acquisition problem invalidated the 2002 WLEF flux data.

4.3.1 Daytime

4.3.1.1 Estimate of CBL height

Based on the measurements from a 915MHz boundary layer profiling radar deployed nearby the WLEF tower in 1998 and 1999, an empirical model relating the fair-weather CBL depth to the integral of the surface buoyancy flux was developed by Yi et al. (2001). In this study, the empirical model is used to estimate the CBL depth in terms of the sensible heat flux measured at the tower.
The characteristic length scale of land cover heterogeneity in the studied region is estimated to be smaller than 4km according to a vegetation map derived from LANDSAT Thematic Mapper (TM) satellite imagery (WiDNR, 1998). Surface heterogeneity will not significantly impact the CBL characteristics if its length scale is smaller than a critical value, estimated to be $0.8U/h/w^*$ with the boundary convective scaling argument (Mahrt, 2000). With typical values of $U=10\text{m/s}$, $h=1000\text{m}$, and $w^*=1\text{m/s}$, the critical length scale is about 8km that is larger than the horizontal length scale of surface heterogeneity. Therefore, the surface heterogeneity in the studied region should not significantly influence the bulk of the CBL and the estimated CBL height is somewhat representative of the region.

4.3.1.2 Estimate of CO$_2$ mixing ratio just above the CBL

CO$_2$ mixing ratio just above the CBL is usually approximated as the baseline mixing ratio (Denmead et al., 1996; Raupach et al., 1992) or the mixing ratio measured at oceanic sites (Levy et al., 1999) when no direct measurements exist. Such approximations, however, can result in significant errors when an air mass has passed over the continent for a long time (Levy et al., 1999). As a result, it is inappropriate to apply those approximations at our site (Helliker et al., 2004; Yi et al., 2004). To a first order approximation, $[c]_+$ can be determined more directly via a combination of CO$_2$ mixing ratio measurements at the tall tower in the lower part of CBL and aircraft profile measurements in the lower part of troposphere (Tans, 1996). When the CBL height is lower than 396m, the top measurement level at WLEF is uses for a measure of $[c]_+$. 
When the CBL is deeper than 396m, CO₂ mixing ratios just above the CBL are predicted with the following equation,

\[
\frac{\partial [c]_a}{\partial t} = \left( \frac{\partial C}{\partial z} \right)_a \left( \frac{\partial h}{\partial t} - w_a \right),
\]  

(4.8)

where two assumptions have been made. First, the effects of the horizontal advection above the CBL are negligible in comparison with those of the vertical advection; Second, the time rate of change of CO₂ mixing ratio above the CBL is small compared with that within the CBL, which is justified by the observed evolution of CO₂ mixing ratio profiles in the CBL and the lower free troposphere (symbols in Figure 4.2). \( (\partial C / \partial z)_a \) is the vertical gradient of CO₂ mixing ratio just above the CBL, which can be formulated to first order as,

\[
\left( \frac{\partial C}{\partial z} \right)_a = \frac{[c]_a - C_H}{h - H},
\]  

(4.9)

by assuming that CO₂ mixing ratios are distributed linearly with height above the CBL between \( h \) and \( H \), where \( H \) is a height in the lower troposphere determined empirically by observed vertical profiles of CO₂ mixing ratio in this area; \( C_H \) is the CO₂ mixing ratio at \( H \), which can be estimated from aircraft data. The analysis of all aircraft profiles indicates that the assumption in Eq. (4.9) is reasonable above the CBL up to a height of 3 or 4 km. Figure 4.2 presents typical vertical profiles of CO₂ mixing ratios from near the surface to the lower troposphere in spring, summer and autumn. The CO₂ mixing ratio generally decreases with height in the dormant season due to the net release of CO₂ by ecosystems, while increasing with height in the growing season due to the net uptake of CO₂ at the surface. The tropospheric profiles are not well-mixed because the exchange of CO₂
between the atmospheric boundary layer and upper troposphere by turbulent transport such as entrainment is not as efficient as mixing within the CBL (Helliker et al., 2004; Hurwitz et al., 2004). In addition to the profiles from the NOAA CMDL aircraft measurements, the linear assumption can be tested more by examining the linear correlation coefficient between the CO₂ mixing ratio and measurement height above the CBL using powered parachute data (Figure 4.3). When the standard deviation of CO₂ mixing ratio in the layer between the CBL top and 4km is smaller than 0.5 ppm, the change of CO₂ mixing ratio with height cannot be resolved due to the limited precision of the instruments. In this case, CO₂ mixing ratio can be approximated as being invariant with height in comparison with the CO₂ jump at the CBL top. Observations showed that CO₂ mixing ratio can be uniformly distributed with height just after a large scale disturbance in the atmosphere such as thunderstorms or frontal passages (e.g., Hurwitz et al., 2004). When the CO₂ mixing ratios change significantly above the CBL, i.e., standard deviation > 0.5 ppm, the assumption of a linear profile can explain more than 80% ($R^2>0.8$, where $R$ is the correlation coefficient) of the change in CO₂ mixing ratio with height for more than half of the cases (Figure 4.3). Most of the correlation coefficients in August are positive, indicating that the CO₂ mixing ratio increases generally with height; the coefficients are either positive or negative in the transition months of the growing and dormant seasons (May and October).

According to the analyses of the vertical profiles of CO₂ mixing ratio from the powered parachute and aircraft measurements, $H$ is taken as 3500m, where the mixing ratio ($C_{H}$) at a given time is estimated by fitting the averaged NOAA aircraft measurements between 3 and 4 km as a function of time. The value of $[c]_+$ estimated
from the tower measurements gives the initial condition for Eq. (4.8) when the CBL height is close to the top level of the tall tower each morning. The results show that the CO₂ jump ([c] - [c]ₘ, [c]ₘ is measured from the tall tower) is negative in the day under fair weather conditions during the dormant season, indicating that the CO₂ mixing ratio is smaller above the CBL than within the CBL. In contrast, the jump is positive during the growing season (solid lines Figure 4.4). During the dormant season, the average jump is about -2 ppm, which does not change significantly with month and with time of day. During the growing season, the jump, however, changes significantly both with month (mostly due to changes in pheneology) and with time of day (in response to changes in photosynthetically-activated radiation (PAR)). In general, the CO₂ jump is larger in the summer months than in the months of late spring and early autumn. For instance, the jump at 1500 LST is about 2 ppm in May, increases to 9 ppm in July, and then decreases to 3 ppm in September. The change in the jump with time of day is the largest in summer. The jump can be as large as negative tens of ppm in the early morning (Yi et al., 2004) (not shown) as a result of the very high concentration of the respired CO₂ accumulated near the surface in summer. With PAR increasing from morning to middle afternoon, CO₂ within the CBL is rapidly assimilated by the photosynthetic uptake of the ecosystem on the surface, leading to a large positive jump in the middle afternoon. In other months during the growing season, the change during the course of a day is smaller mainly because of weaker assimilation and respiration rates.
4.3.1.3 Entrainment flux and vertical advection term

The entrainment flux, characterizing the net exchange of CO₂ across the CBL top, can be estimated using the zero-order jump model as,

\[ F_e = -\Delta C \frac{\partial h}{\partial t} + w_i \Delta C, \quad (4.10) \]

where \( \Delta C = [c] - [c]_m \); the first and second terms on the RHS of the above equation represent the exchange of CO₂ due to the varying CBL top and due to the vertical advection, respectively. Because the budget method is applied during the growing period of the CBL when its depth is reasonably predictable, the time derivative of \( h \) is non-negative and reaches its maximum value in the middle morning and becomes smaller in the afternoon, indicating that the diurnal pattern of the time derivative of \( h \) differs from that of the CO₂ jump. Thus, the first term on the RHS of Eq. (4.10) reaches the maximum value in the middle or late morning during the growing season. The pattern of the diurnal change in the CBL depth does not vary significantly with season compared with that of the CO₂ jump. As a result, the seasonal change in the diurnal cycle of the term is determined primarily by that of the CO₂ jump. This term is about -2 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) during the period from 10:00 to 15:00 LST in May, decreases to the most negative value of about -8 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in June, and then increases to about -3 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in the last month of the growing season. In contrast, the value of this term does not significantly vary with the month during the dormant season, remaining at about 2 \( \mu \text{mol m}^{-2}\text{s}^{-1} \).

The sign of the vertical advection term is opposite to that of the CO₂ jump as a result of the negative vertical velocity under the atmospheric conditions being considered.
Like the first term, the behavior of the monthly and diurnal changes in this term is determined mainly by the CO$_2$ jump because the vertical velocity does not significantly vary diurnally and monthly compared with the CO$_2$ jump. The vertical advection term, in general, has the same magnitude as that of the entrainment flux component due to the growing CBL top. The average value of the vertical advection term from 10:00 to 15:00 LST is as large as about -5 µmol m$^{-2}$s$^{-1}$ in summer months, while it is about 0.5 µmol m$^{-2}$s$^{-1}$ during the dormant season. The magnitude of the vertical advection term is about 20-30% of the magnitude of the surface fluxes.

The monthly diurnal-average of the total entrainment flux ($F_e$) is presented in Figure 4.5. In general, the magnitude of the entrainment flux from 10:00 to 15:00 during the growing season is about 50% of the surface flux in May and September and increases to 80% in summer months. In addition, unlike the two individual terms contributing to the entrainment flux in Eq. (4.10), $F_e$ does not vary significantly from late morning to middle afternoon. In the early morning, the entrainment flux value is larger and sometimes has the opposite sign to the surface flux due to the negative CO$_2$ jump (Figure 4.4 and Figure 4.5c, f and g). During the dormant season, the entrainment flux has similar magnitude but opposite sign to the time rate of change term (see the next section). Therefore, uncertainty in either of the two terms may result in significant relative errors in the surface flux estimate in the dormant season. This suggests that both terms are significant in estimating the surface flux using the budget method.
4.3.1.4 Time rate of change term

The evolution of the mean mixing ratio of CO$_2$ in the growing CBL depends primarily on the CBL depth, entrainment flux, and surface flux. The contribution of the surface flux to the change in the mean mixing ratio can be diluted by the increasing CBL depth, and sometimes can be cancelled out totally or partially by the entrainment flux. As a result, the time rate of change in the mean mixing ratio in the day is usually small compared with that at night even though the magnitude of the daytime surface flux is larger, as shown by observations (Yi et al., 2001).

During the developing period of the CBL in the growing season, the time rate of change term is in general less negative in the early morning than in the late morning or afternoon (Figure 4.5). On average, this term is negative regardless of month at this site, indicating that the mean CO$_2$ mixing ratio decreases with time as the CBL grows. The average value of this term from 10:00 to 15:00 LST is about -4 µmol m$^{-2}$s$^{-1}$ in the growing season, which is about 20% of the surface flux in June, July, and August and 50% in May and September. It is about -2.5 µmol m$^{-2}$s$^{-1}$ in the dormant season, larger in magnitude than the surface flux. The signs of the mixing ratio time rate of change and the surface flux are the same during the growing season, but opposite during the dormant season.

The comparison also indicates that the time change rate term is only slightly more negative in the growing season, in spite of much stronger CO$_2$ uptake, than in the dormant season, suggesting the important roles of the mixing within the CBL and entrainment at the CBL top in estimating surface flux with the budget method.
4.3.1.5 Regional fluxes and Comparisons with measurements at three levels of the WLEF tower

On average, the inferred regional surface flux is about 2 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) during the daytime in the dormant season as shown in Figure 4.5 (a, b, and h), indicating a small net release of CO\(_2\) in the region. The surface flux is smaller, sometimes slightly negative, in the middle of the day as a result of photosynthetic activities of coniferous forests in the region. This effect is also seen in fluxes measured at 396m of the WLEF tower using the eddy-covariance method (circles in Figure 4.5). In contrast, the surface flux is significantly negative in the middle of the day during the growing season with the largest net uptake of CO\(_2\) being found in June and July. The magnitude of the average surface flux from 10:00 LST to 15:00 LST is about 7 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in May, increases to 18 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in July, and then decreases to 10 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in September. During the course of a day, the magnitude of the surface flux is generally larger in the late morning and early afternoon than that in the early morning (e.g., 08:00LST) and middle afternoon (e.g., 15:00LST). The monthly and diurnal variations of the estimated surface flux are comparable to those measured using the eddy-covariance method. The magnitude of the regional surface flux is generally larger than that measured at 396m of the tower in the growing season, probably due to different footprint areas in addition to uncertainties of both eddy-covariance and ABL budget methods.

Further comparison with tall tower measurements is shown in Figure 4.6. The monthly diurnally-averaged NEE values measured at 30, 122, and 396m are overall about 40%, 58%, and 71% of those inferred from the CBL budget method, respectively,
indicating that NEE values measured at the tower are closer to regional estimates of NEE as the measurement height increases. This phenomenon can be explained partially by footprint modeling results. Figure 2.8 indicates that the footprint area increases with the measurement height. Figure 2.9 implies that measurements at the two higher levels could be more representative of the region than those at the lowest level in terms of fractional weights of the ecosystem types to the total measured flux in their footprint areas in comparison with the fractional coverage of the respective ecosystem types in the region being studied. The comparison also suggests that daytime measurements at the lowest level of the tower likely underestimate regional NEE.

4.3.1.6 Fitting regional-scale surface fluxes

To estimate the integrated surface fluxes over seasons and reduce random errors, the inferred regional surface fluxes are fitted to ecosystem models that formulate CO₂ flux as a function of environmental variables such as air temperature and PAR that can be representative of the region to some extent. Figure 4.7 presents the bin-averaged NEE estimated using the CBL budget method, together with those measured at the WC, LC, and WLEF towers, as a function of PAR in each month, where the inferred NEE values are binned by PAR in increments of 200 µmol m⁻² s⁻¹. The mean and its standard deviation are calculated for each bin. Because the CBL budget method is applied in midday under fair weather conditions, the resulting surface fluxes are available only in the range of large PAR values. In general, during the growing season, NEE measured at the WC tower is the largest for a given PAR among those measured at the three towers with exception
of the month of May (when leaves are just starting to grow), while NEE measured at the LC tower is the smallest. The regional estimates of the surface flux are intermediate, which is overall closer to those measured at the highest level of the WLEF tower and larger than those measured at other two levels. Inferred regional surface fluxes are larger than measured WLEF fluxes in particular during the middle of the growing season. The comparison between regional fluxes and tower measurements also suggests that the tower measurements can give reasonable bounds for the regional NEE estimates at a given PAR. Provided that this is true over the full range of PAR, the mean or median value of NEE measured at the towers might be a reasonable estimate of the regional NEE under low PAR conditions because the differences in NEE among the tower measurements generally decrease as PAR decreases (Figure 4.7). Under this assumption, the light response curve of the ecosystem during the growing season used to describe the response of regional CO₂ flux to PAR can be modeled with a rectangular hyperbola, i.e.,

\[ NEE = \frac{\alpha I P_m}{\alpha I + P_m} + R, \]  

or a linear equation,

\[ NEE = \alpha I + R, \]  

where \( I \) is PAR (\( \mu \text{mol quanta m}^{-2}\text{s}^{-1} \)); \( P_m \) (\( \mu \text{molC m}^{-2}\text{s}^{-1} \)) is the assimilation rate at saturating PAR; \( \alpha \) is the apparent quantum yield (\( \mu \text{molC/\mu mol quanta} \)), and \( R \) is the dark respiration rate (\( \mu \text{molC m}^{-2}\text{s}^{-1} \)). Eq. (4.12) is a special case of Eq. (4.11) with \( P_m \) being infinite, which was observed for some types of ecosystems (Ruimy et al., 1995). Table 4.1 summarizes the parameters of the model fitted to the regional NEE estimates,
and fitted to the NEE measured at the towers for comparison. More than 70% of the change in the average regional flux with PAR can be explained by the fitted curves. Overall, the magnitude of the apparent quantum yield on the regional scale is smaller than those fitted from the tower measurements. The magnitude of the regional saturated assimilation rate \( P_m \) is smaller than but closer to the largest value fitted among the three towers. The regional respiration rate \( R \) is bounded by those inferred from the tower measurements with the WLEF and LC sites having the largest and smallest values, respectively. It is also noticed that both regional parameters \( P_m \) and \( R \), corresponding to fluxes at infinite and zero PAR, respectively, may not be equal to the averages of those inferred from tower measurements weighted by the area of the respective ecosystem types that the sites can represent. Reasons for the inequality are not clear. One likely explanation, other than measurement errors and non-representative sampling at towers, is that the eddy-covariance method cannot measure the fluxes on intermediate scales between 1km\(^2\) and 1000km\(^2\), while the budget method does integrate fluxes over these scales from all ecosystem types including those that are not represented by current measurements at the towers. This is a key strength of the top-down method.

Comparison of the fitted parameters also indicates that the relationship between the CO\(_2\) flux and PAR on the regional scale is more linear than that on the scale of the tower measurements, which can be seen from Figure 4.7 or quantitatively characterized by the difference in the coefficients of determination \( (r^2) \) between the linear and the rectangular hyperbola models fitted to the same data set. Previous research also found that scaling up in time and space tends to linearize the light response curve and to decrease the apparent quantum yield. This was determined by comparing the flux versus
PAR relationships fitted to data from three methods on increasing spatial scales, i.e., enclosure, ground-based micrometeorological measurements, and aircraft-based flux measurements (Ruimy et al., 1995). This tendency is consistent with and verified further by that indicated in this study.

The response of the regional NEE to PAR, however, cannot be well described by the light response curve (Figure 4.7a, b, and h) during the dormant season as a result of significantly reduced photosynthetic activities. Therefore, attempts are made to examine the relationship of NEE and air temperature as shown in Figure 4.8, which can be modeled with an exponential equation, i.e.,

\[ \text{NEE} = a e^{bT}, \]  

(4.13)

where \( T \) is the air temperature at 30m in °C above the ground; \( a \) is NEE at \( T=0 \) °C; and \( b \) is the coefficient describing the sensitivity of NEE to the change in temperature. The equation is identical mathematically to the widely-used Q-10 model (Lloyd and Taylor, 1994). This model describes the response of NEE to temperature at the WC and LC sites very well in winter months from November through April with \( r^2 \) greater than 0.9. Only about 30% of the variation of the regional estimates of NEE with air temperature, however, can be explained by the exponential model. Nevertheless, Figure 4.8 still shows that the regional NEE during the dormant season is generally bounded by the measurements at towers at a given air temperature.

The responses of NEE to environmental conditions can be described by the exponential model from November to April except that the measurements at the 30m of the WLEF are best fitted by the light response model in terms of \( r^2 \) (Table 4.1). The
different selections for the best model during the dormant season are primarily because the footprint areas are covered in part by coniferous forests at the WLEF site, while being covered mostly by deciduous forests at the other sites with little photosynthetic activities in the season. On the larger scale, positive daytime NEE values indicate that the amount of CO$_2$ respired is more than that assimilated by the ecosystem during the dormant season. But neither of the processes is overwhelmingly dominated, accounting partially for the difficulty in describing the regional NEE alone either with the light response curve or with the exponential model. In contrast, at tower sites, particularly WC and LC sites, the respiratory process dominates during the dormant season except for the transition months.

4.3.1.7 Monthly averaged daytime NEE and comparison with tower measurements

With the above fitted models and measured temperature and PAR at the WLEF tower, monthly-averaged daytime regional NEE values are calculated during the growing season, while seasonally-averaged NEE is calculated during the dormant season because not much data is available. The same calculations are made for NEE at the towers for comparison.

The NEE value estimated from the budget method is about 0.91±0.10 µmol m$^{-2}$s$^{-1}$ in the day during the dormant season, suggesting that the ecosystem in the region is a small net source of CO$_2$. The regional estimate is larger than those measured at the LC tower and 30 and 122m of the WLEF tower, while the difference among NEE values
estimated from the budget method, observed at the WC tower and 396m of the WLEF tower is not statistically significant. The region, however, is observed to be a significant sink in the day during the growing season with the maximum net uptake (-7.92±0.54 µmol m⁻²s⁻¹) occurring in June (Figure 4.9). The daytime regional NEE is closer to that measured at 396m of the WLEF tower in June, July, and August. In May, the daytime regional NEE (-2.57±0.22µmol m⁻²s⁻¹) is closer to those measured at 30m and 122m of the WLEF tower. In September, the daytime regional NEE (-4.57±0.46 µmol m⁻²s⁻¹) is in between those measured at the Willow Creek tower and at 396m of the WLEF tower. The magnitude of NEE measured at the WC tower is the largest during the growing season except early spring (May) when leaves are just starting to grow. The smallest magnitude of NEE is observed at the LC tower in May and June and at 30m of the WLEF tower in other months during the growing season. Although the regional NEE is bounded by the existing tower measurements, the tower measurements are insufficient to constrain the regional NEE because the difference between the maximum and minimum NEE values at the towers is as large as about 7 µmol m⁻²s⁻¹ in summer, which is almost close to or even larger than the magnitude of the regional flux itself. The large difference is due to varying responses of the ecosystems to similar environmental conditions over the region, implying the importance of measuring NEE for all ecosystem types. Comparisons in further detail are beyond the scope of this chapter.
4.3.2 Nighttime

One benefit of applying the NBL budget method is that the evolution of the mean mixing ratio in the NBL with time is larger than that in the daytime CBL, and hence more easily detected (Denmead et al., 1999; Denmead et al., 1996). But regional NEE is still more difficult to estimate using the budget method at night than in the day at least for three reasons.

First, it is necessary to measure the vertical profiles of CO₂ mixing ratio with high resolution in the NBL because of the large vertical gradients unlike those in the daytime CBL where measurements at one level can approximate the mean CO₂ mixing ratio in the CBL either directly if the level is above the surface layer (Yi et al., 2001), or indirectly with the similarity theory (Denmead et al., 1999) if the level is within the surface layer. The highest level of the vertical profile should be equal to or higher than that of the NBL top above which the vertical turbulent flux can be negligible; otherwise, the vertical turbulent flux at the top level of the profile needs to be considered. Practically, not many sites have such profile measurements except in special campaigns (e.g., Eugster and Siegrist, 2000; Pattey et al., 2002).

Second, the depth of the nocturnal boundary layer is difficult to predict. It is hard to apply the budget method, particularly when the radiative inversion layer is deep or absent. One way to avoid directly predicting the NBL depth is to measure the vertical profile of CO₂ mixing ratio to a level that is higher than the typical NBL top. In this regard, we can take advantage of the measurements at the tall tower because the NBL depth is usually smaller than the highest level of the measurements (396m) in many
cases. The cases when the NBL top is higher than 396m should be identified and removed to avoid possible systematic errors.

Finally, respired CO$_2$ may not be well-mixed in the atmosphere under very stable stratification, leading to less representative regional fluxes and possible systematic errors resulting from the use of vertical profiles measured at a single location. Therefore, the fluxes calculated under these conditions should be screened out.

Three steps are made to infer the surface fluxes from the vertical profiles of CO$_2$ mixing ratio measured at the 447m-tall tower. The one-dimensional budget equation of CO$_2$ mixing ratio (Eq. 4.7) is first integrated from surface to the highest measurement level for all nights when data are available. The results are called first-guess NEE. Then the first-guess NEE values are screened to eliminate systematic errors and possible non-representative sampling under certain meteorological conditions; the screening criteria are developed by comparing the first-guess NEE against atmospheric stability parameters. Finally, the screened results are approximated as the nighttime NEE and used to fit the exponential model. This section starts with identifying screening criteria.

4.3.2.1 Screening Criteria

Under near-neutral and very stable conditions, systematic errors may be significant in the first-guess NEE. The NBL top may be higher than the highest level of the measurements under weak stable or nearly-neutral conditions, possibly resulting in biased estimates (typically underestimates) of NEE. In this case, the vertical turbulent flux at 396m of the tower can be significantly nonzero, hence integrating the budget
equation from surface only to 396m may result in underestimation of the surface flux. Under very stable conditions where the NBL is shallow, the first-guess NEE, however, either may be underestimated, or may be less representative of the region, or both. The CO₂ emitted from ecosystems is usually contained in a shallow air layer near the surface where the mixing ratio varies noticeably during night, occurring often with the presence of a near-surface inversion. In this case, the vertical mixing is restricted, causing the sub-canopy to be decoupled from overlying air (Mahrt et al., 2000). Thus, the variability in CO₂ mixing ratios near the surface, which is much larger than the variability at higher levels, cannot be totally detected by the single sub-canopy observation at the lowest level (11m) of the tower, resulting in errors. In addition, using the vertical profiles measured at a single location ignores the heterogeneity in sub-canopy source distribution as well as drainage flows that form readily near the surface even with a small slope of terrain. These flows are believed to account for a significant part of the errors of the NEE estimates. From another perspective of the representativeness of sampling, the spatial distribution of CO₂ mixing ratio may become more locally variable with increasing stability due to heterogeneous terrain, or canopy, or both. As a result, the measurements at a single location might be less representative of the region under very stable conditions.

According to the above analyses, systematic errors can occur both under very stable conditions and under very weak or neutral conditions, which should be screened out. The screening criteria are developed by examining the NEE values as a function of two micrometeorological stability variables: the friction velocity and Monin-Obukov length. Both micrometeorological variables can be estimated by surface measurements at towers.
4.3.2.1.1 Criteria under weak stable or near-neutral conditions

The friction velocity, an indicator of turbulent strength, is usually used to express screening criteria in the literature. The first-guess NEE are segregated by friction velocity ($u^*$) during the growing season and used to identify a threshold of $u^*$ above which this tower-based budget method systematically underestimates NEE. The underlying principle used for defining the criteria is similar to that for screening nighttime NEE measured at towers (Aubinet et al., 2000; Cook et al., 2004). Ecosystem respiration at night is controlled mainly by air and soil temperature, and, sometimes, soil moisture. It should be independent of the level of turbulence described by $u^*$ on canopy scales if the correlation between $u^*$ and temperature is insignificant or is removed. To remove the impacts of the changes in soil and air temperatures on the comparison of NEE at different $u^*$ values, the NEE values are normalized by the simulated NEE with the temperature-dependent model (i.e., Eq. 4.13) fitted to the first-guess NEE results (Goulden et al., 1996). The first-guess and normalized NEE values drop systematically with increasing $u^*$ when $u^*$ is larger than 0.65 ms$^{-1}$, as indicated in Figure 4.10. With the same normalized approach, Figure 4.11 presents the NEE values against the Monin-Obukov length, indicating that the systematical dropping of NEE values when $L$ is greater than approximately 400m. Based on the above two thresholds, about 10% of the first-guess NEE is discarded.

Applying the above criteria is expected to remove the cases where the vertical turbulent fluxes are non-zero at the top level of the measurements at the tower. To examine this hypothesis, Figure 4.12 presents the vertical turbulent fluxes of CO$_2$ directly measured at 30m, 122m, and 396m of the tall tower during the growing season as a
function of the stability parameters. The vertical turbulent fluxes vanish at 30, 122, and 396m when the $u^*$ values are smaller than 0.025, 0.2, and 0.65 ms$^{-1}$, respectively (Figure 4.12a), or when the $L$ values are smaller than 0.5, 30, and 500m, respectively (Figure 4.12b). The thresholds of $u^*$ and $L$ increase with height, indicating that CO$_2$ released from the surface is contained within a thicker layer under less stable atmospheric conditions. More importantly, the thresholds of $u^*$ and $L$ implied from the eddy-covariance fluxes at 396m are in good agreement with those derived from the analyses of the first-guess fluxes based only on the mixing ratio measurements; this in turn indicates that the methodology used here works well to find thresholds to screen the cases when the vertical turbulent flux is non-zero at the top level of the measurements.

4.3.2.1.2 Criteria under very stable conditions

The first-guess NEE values only drop slightly under weak turbulent conditions compared with those under near-neutral conditions, which distances from our expectations according to physical analyses and observation for unclear reasons. As a result, this examination method does not work well for identifying thresholds under very stable conditions. Alternatively, the criteria under very stable conditions are determined based on the following aspects.

First, the analyses of the eddy-covariance measurements at the towers in the region provide evidence that NEE is underestimated under very stable conditions mostly due to difficulties in considering the effects of heterogeneity and drainage flow beneath the canopy only from measurements at a single location. Under very stable conditions,
NEE could be underestimated using the eddy-covariance method by a factor of about 2 at WLEF (Ricciuto et al., submitted) and WC (Cook et al., 2004) sites when the friction velocity is smaller than 0.2 and 0.3 m s\(^{-1}\), respectively. In other words, about half of NEE cannot be detected by the measurements under very stable conditions. This failure is not only due to possible underestimation of turbulent fluxes at low \(u^*\) values (Goulden et al., 1996), but also due to lack of measurements of spatial distribution of CO\(_2\) mixing ratio, particularly within the canopy. In this case, NEE values inferred from the mixing ratio measurements at a single tower are questionable. As a result, the first-guess regional NEE values under low \(u^*\) conditions are discarded to remove possible systematic errors and to estimate NEE with more confidence.

Second, measurements at a single location are affected primarily by local sources under very stable conditions, and hence are less representative of a larger area. This is a result of restricted vertical mixing and the resulting patchy distribution of CO\(_2\) mixing ratio. For example, the integral of the time-derivative of CO\(_2\) mixing ratio from the surface to 30m at the WLEF site, i.e., Eq. (4.7) with \(h_i\) being equal to 30m, corresponding to part of the surface flux, is significantly larger than that at the WC site, on average, by about 15\% of NEE under low \(u^*\) (\(\leq 0.2\) m/s) conditions (Figure 4.13). In contrast, the average magnitude of the relative difference of the integral is only about 2\% when \(u^*>0.2\) m/s, much smaller than that under very stable conditions. The comparison, therefore, suggests that NEE values calculated under very stable conditions need to be discarded from the perspective of representative sampling.

In consideration of the above evidence and the number of data points screened, a \(u^*\) threshold of 0.2m/s is selected in the following evaluation, below which either
estimated NEE values are less representative, or the effects of the possible drainage flow and heterogeneity beneath the canopy are significant but not detectable, or both. About 30% of the first-guess NEE values are discarded on this basis. The large percentage implies that these conditions occur frequently, and significant errors could occur if the calculation were not screened.

The extent of the surface flux which the NBL budget can represent can be estimated as the product of the characteristic horizontal wind speed and time scale. For the well-mixed conditions in the calculation, i.e., \( u^* \) is between 0.2 and 0.65 m/s, the horizontal scale of the inferred NEE is about 7 to 18 km with time scale of 1 hour and wind speed of 2-5 m/s.

### 4.3.2.2 Model fitting and parameters

The exponential model, Eq. (4.13), is used to fit the screened nighttime NEE. The estimated parameters are listed in Table 4.2. The fitted NEE values from the budget method are bounded by the tower measurements (Figure 4.14). During the growing season, about 80% of the change of the inferred regional NEE with air temperature can be explained by the fitted model. During the dormant season, the inferred NEE values are more scattered, with the determination coefficient being about 50%. The temperature-dependent nighttime respiration rates measured at all towers with exception to the LC tower can be well described by the exponential model with the \( r^2 \) values ranging from 0.8 to 0.95. Preliminary results show that the respiration rate at the LC site is related to the
depth of water table of the wetland, and the temperature-dependent model is insufficient to describe this behavior.

During the dormant season, the largest and smallest respiration rates at 10°C, denoted as $R_{10}$, are observed at the WC tower and at LC tower, respectively; $R_{10}$ inferred by the budget method is the second smallest. During the growing season, $R_{10}$ measured at the WLEF tower and the LC tower are the largest and smallest, respectively, with the respiration rate inferred by the budget method being the second largest (Table 4.2). $R_{10}$ values are larger in the growing season than in the dormant season, due to phenological changes. All $Q_{10}$ values observed at the towers and inferred by the budget method are larger, implying that the respiration rates for the respective ecosystems are more sensitive to temperature change during the dormant season than during the growing season. The $Q_{10}$ value inferred from the ABL budget method is the largest in both seasons, indicating that the respiration rate of the regional ecosystem is the most sensitive to temperature change.

4.3.2.3 Seasonally-averaged regional nighttime fluxes and comparison with tower measurements

The averaged nighttime NEE derived from the budget method is about 3.89±0.19 µmol m$^{-2}$s$^{-1}$ during the growing season, which is between the NEE values measured at the WC tower (3.24±0.10µmol m$^{-2}$s$^{-1}$) and at the WLEF tower (5.27±0.20µmol m$^{-2}$s$^{-1}$) (Figure 4.16). Among the three tower measurements, the nighttime NEE is the smallest at
the LC tower (1.61±0.35 µmol m^{-2}s^{-1}) with the largest NEE being measured at the WLEF tower. The comparison also indicates that measurements at the WLEF tower likely overestimate regional nighttime NEE. During the dormant season, the nighttime NEE measured at the LC tower is still the smallest, with the WC tower likely measuring the largest NEE. The regional NEE (0.72±0.07 µmol m^{-2}s^{-1}) is likely between those measured at the WLEF tower (0.89±0.10 µmol m^{-2}s^{-1}) and at the LC tower (0.37±0.04 µmol m^{-2}s^{-1}). The differences in NEE values measured at the towers and inferred from the budget method are small during the dormant season in contrast to those during the growing season.

4.3.3 Discussions

Uncertainties due to the assumptions used in the budget calculation are discussed in this section.

4.3.3.1 Uncertainty due to uses of measurements at the single location

The vertical profiles of CO_{2} mixing ratio measured at a single location, which are assumed implicitly to be representative of the region, are usually used to calculate the time rate of change term and the entrainment flux term. Possible uncertainties resulting from this assumption are assessed below.
During the nighttime, the CO$_2$ mixing ratio changes noticeably with time, particularly near the surface due to surface fluxes. The time rate of change term, a major component in the budget equation, is usually approximated as the mean nighttime surface flux. As described in section 4.3.2.1.2, the difference in this term below 30m observed at the WLEF and WC towers (Figure 4.13) is significant under low values of $u^*$, while small under well mixed conditions. To compare the time rate of change term at higher levels, we compute the derivative of CO$_2$ mixing ratio with respect to time measured at 76m of the WLEF tower, Brule tower (46.466833N, 91.566743W), and Fence tower (45.735359N, 88.427305W) under fair weather conditions in August 2003, respectively. The latter two towers are located about 100km northwest and southeast of the WLEF tower, respectively. Measuring CO$_2$ mixing ratio with high precision at these towers is a part of a project aimed at inferring regional fluxes (for details, see http://rflux.psu.edu).

At night, the monthly averaged deviation of the derivative at the WLEF tower from the average of the respective derivatives calculated at the three towers is about 0.001 $\mu$ molC m$^{-3}$s$^{-1}$ (filled circles in Figure 4.15). Unlike those below 30m, the averaged deviation value is not strongly dependent on turbulent conditions, though the deviation is more scattered in the case of lower $u^*$ values. By assuming that the depth of NBL is of order of the 400m, the error in estimating surface fluxes would be of the order of about 0.3 $\mu$ mol m$^{-2}$s$^{-1}$ due to the sampling of the change in CO$_2$ mixing ratio at high levels (e.g., 100m to 400m) at the single location (WLEF tower), which is about 5-10% of the surface flux. With the error of sampling near the surface (<100m) being counted, the total error would be about 30-35% and 9-14% under the conditions of low ($\leq$0.2 ms$^{-1}$) and high (>0.2 ms$^{-1}$) $u^*$ values, respectively; here the error due to sampling between surface and 100m is
estimated by multiplying half of the relative difference of the integrated time derivative term below 30m shown in Figure 4.13 by a factor of 3.3 (=100m/30m).

In the daytime, the measurement of CO₂ mixing ratio at 76m can be approximated as the average of mixing ratio in the CBL according to observation (Yi et al., 2001). The mean deviation of the time derivative term from the mean is about 0.0016 µmol m⁻³ s⁻¹ in August (open squares in Figure 4.15). With a typical CBL depth of 1000m, the monthly averaged error due to sampling at the single point is about 1.6 µmol m⁻² s⁻¹, which is about 10% of the surface flux.

With respect to the effects of single-point measurements on estimating the entrainment flux, we can only make a crude evaluation because observational data of the horizontal gradients of CO₂ mixing ratio are unavailable. Provided that the difference of CO₂ mixing ratio is about 1 to 2 µmol mol⁻¹ over a horizontal distance of 10² km under fair weather conditions during the growing season and that the horizontal gradient of CO₂ mixing ratio above the CBL is negligible compared to that within the CBL, the error in the entrainment flux of CO₂ would be about 0.4 to 0.7 µmolC m⁻² s⁻¹ in the afternoon when the growth rate of the CBL depth is negligible and with a typical mean vertical velocity of the order of 0.01m/s. This error in entrainment flux is negligible compared with the typical surface flux values of 15 µmol m⁻² s⁻¹ in the season. During the growing period of the CBL, the error, however, would be larger due to the large entrainment velocity. For instance, with an increase of 200m in CBL depth per hour occurring in the morning, the error would be about 2 to 4 µmol m⁻² s⁻¹, which is about 10 to 20% of the surface flux.
4.3.3.2 Effects of horizontal advection

The likely ranges of the effects of the horizontal advection on the surface flux estimates due to the heterogeneous distribution of the surface vegetation is evaluated in eight wind direction sectors using the water vapor constraint approach as described in Chapter 5. The overall magnitude of the effects is smaller than about 15-20% of the surface flux in each wind direction in the middle day (from noon to 15:00LST) during the growing season. When winds are from the NE, E, and SE directions, the horizontal advective fluxes are likely positive, while they are likely negative when the winds are from the SW, W, and NW directions. In other wind directions, the signs of the horizontal advective fluxes cannot be determined. The monthly averaged impact of horizontal advection on the surface flux estimates using the one-dimensional budget equation in this study range from -1.7 to 0.64 µmol m\(^{-2}\)s\(^{-1}\), from -2.06 to 0.30 µmol m\(^{-2}\)s\(^{-1}\), from -2.36 to 0.33 µmol m\(^{-2}\)s\(^{-1}\), from -1.8 to 0.68 µmol m\(^{-2}\)s\(^{-1}\), and from -2.5 to -0.02 µmol m\(^{-2}\)s\(^{-1}\) in months from May through September, respectively. In the evaluation, the seasonal mean contribution of the horizontal advection in each wind direction is used for each month. The overall estimate indicates that the effects of the horizontal advection on the current flux calculation are smaller than 10-15% of the surface flux during the growing season. For a more accurate evaluation, direct measurements of horizontal advection, which are difficult currently to obtain, are needed eventually to eliminate the uncertainty due to ignoring the horizontal advection terms.
4.3.3.3 Effects of the interfacial layer

The budget calculation in this study ignores the effects of the interfacial layer at the top of CBL, which is about 20% of the mean depth of the CBL as indicated by recent observations (Davis et al., 1997). Appendix C evaluates the differences of all terms in Eq. (4.6) for the cases with and without the interfacial layer being considered. In the evaluation, the depth of the interfacial layer is estimated in two ways. One is to assume that the depth is a constant fraction (e.g., 20% in this calculation) of the CBL depth. The other is to predict the depth with a model that is fitted to observation as a function of environmental variables (Appendix C). CO₂ mixing ratios are also assumed to be distributed linearly with height in the interfacial layer in order to facilitate the calculation. Depending on whether or not the interfacial layer is considered, the changes in each of the time rate of change term and entrainment flux term is about 10% of the surface flux, increasing with the depth of the interfacial layer. However, the changes in both terms are generally inversely correlated, leading to a small change (≤1 to 5% of the surface flux) in the surface flux estimate that is the sum of the two terms.

4.3.3.4 Other sources of uncertainty and future work

In the CBL budget calculation, uncertainty in predicting the CBL depth, the vertical velocity at the CBL top, and the CO₂ jump could be other sources of errors because those variables are used to estimate the entrainment flux that is more than 50% of the surface flux in magnitude according to the calculation in this study. The CBL depth
is estimated from the model that actually predicts the mean depth given the surface sensible flux. Thus, it is inappropriate to interpret the surface flux estimated on a short temporal scale such as hours, though the time rate of change term is calculated hourly. Instead, the results should be interpreted on long time scales. The comparison of the monthly diurnally-averaged NEE inferred from the CBL budget method with the tower measurements indicates that the CBL budget method produces reasonable patterns of the diurnal change and magnitudes of the surface flux. The CO₂ mixing ratio above the CBL is estimated by interpolating the direct measurements within the CBL and in the lower troposphere based on the assumptions that can be justified by observations, which predicts reasonable seasonal and diurnal variations of CO₂ jump. Therefore, significant errors in the monthly mean CO₂ jump would not be expected. Under fair weather conditions, the large-scale vertical velocity is more likely negative. Therefore, the long term averaging cannot eliminate or reduce the contribution of the vertical advection term unlike the horizontal advection term. One difficulty is how to estimate the vertical velocity, which is too small to be measured accurately. Although the vertical velocity from the reanalysis data is reasonable in the order of magnitude, the uncertainty due to using these data in the budget calculation is unknown. More elaborate approaches such as aircraft and powered parachute platforms that can measure the variables at the top of CBL would be expected to yield more precise estimates of NEE due to the significant role of the entrainment flux in estimating the surface flux. The requirement for the measurements at the CBL top is a major weakness of applying the CBL budget method particularly for a long time period. Without such measurements, assumptions and empirical models are inevitable. Despite these uncertainties and restrictions, the CBL
budget method still can produce reasonable seasonal and diurnal cycles of regional NEE, which is superior to the estimates that only provide order of magnitude (Denmead et al., 1999). For a region with heterogeneous distributions of vegetation, measurements at more than one location are needed, e.g., the regional flux project going on in Northern Wisconsin (http://rflux.psu.edu).

In the NBL budget calculation, ignoring the advection terms could also be a source of systematic errors in the estimates of nighttime NEE in addition to those previously identified. The estimated NEE could represent a larger extent of the surface under well-mixed conditions than under very stable conditions, but possibly suffer more significant impacts of the advection as a result of the larger wind speed. Attempts have been made to include the vertical advection term in the calculation by extrapolating the reanalyzed vertical velocity from the RUC model to the six levels at the WLEF tower. The resulting fluxes are not reasonable probably due to the combination of the questionable extrapolated vertical velocity data with the large gradient of CO2 mixing ratio in the lower part of the NBL near the surface. With respect to the effects of the horizontal advection on the estimates of nighttime NEE, it is difficult to quantify without measurements. It is, however, believed that the effects of advection can be reduced after long-term averaging due to cancellation from opposite directions. Therefore, the uncertainty in the NEE estimates on the hourly scale is expected to be larger than that on the seasonal scale. To reduce or quantify such uncertainty, the vertical profiles measured at more than one location are needed.
4.4 Summary

By taking advantage of the long term measurements of CO$_2$ mixing ratio profile at the 447-m tall WELF tower, seasonally-averaged NEE values are estimated over the region with the area of 40×40km$^2$ centered at the tower. The annual cumulative NEE averaged in 2000 and 2003 are about -147±33 and -197±34 gC m$^{-2}$, respectively; this indicates that the region is a net sink of CO$_2$. The regional estimate of NEE is between those measured at Willow Creek tower and the WLEF tower derived from the eddy-covariance method. During the growing season, the averaged daytime regional estimate of NEE is between those measured at the WLEF and Willow Creek towers with the largest net uptake being observed at the WC site. The regional estimate of NEE is closest to that measured at 396m of the WLEF tower among the measurements at the three towers. The nighttime regional NEE is also between those measured at the WLEF and WC towers with the WLEF tower observing the largest respiratory flux. During the dormant season, differences in NEE values estimated from the budget method and measured at towers are small. The daytime regional estimate of NEE is bounded by the tower measurements at a given PAR and at a given air temperature during the growing season and dormant season, respectively, while the nighttime regional NEE is bounded by the tower measurements at a given air temperature during both seasons.

In the daytime CBL budget calculation, the CO$_2$ mixing ratio is assumed to be distributed linearly with height above the CBL, which is supported partially by observations. Based on this assumption, the CO$_2$ mixing ratio just above the CBL top is estimated based on the evolution rate of the CBL with time and the direct measurements
of CO$_2$ mixing ratio within the CBL and in the lower troposphere, giving reasonable diurnal and seasonal changes in the CO$_2$ jump across the CBL top. The CO$_2$ mixing ratio is assumed to be constant in the mixed layer, represented by the 396m measurement at the WLEF tower. The CBL depth is predicted with an empirical model that was developed using radar measurements at the site. The vertical velocity at the CBL top is estimated by interpolating those from the RUC reanalysis data to the level of the CBL top. During the growing season, the contribution of the time rate of change term to the surface flux estimate is smaller than that of the entrainment flux term; while both terms are similar during the dormant season. The regional estimates of NEE are fitted with the light response curve during the growing season. The fitted function is more linear than those fitted to data measured at the individual towers.

Nighttime NEE is estimated by integrating the vertical profiles of CO$_2$ mixing ratio from the surface to the highest measurement level of the tower. The cases under very stable ($u^*<0.2$ms$^{-1}$) and near-neutral conditions ($u^*>0.6$ms$^{-1}$), where systematic errors could occur, are screened out in the calculation. NEE estimates that pass $u^*$ screening could suffer the impacts of the advection, which cannot be quantified currently. It would be expected that the impacts of the advection can be somewhat reduced after long-term averaging. The results show that the seasonally-averaged nighttime NEE values are comparable to those measured at the towers.

In the CBL budget calculation, uncertainties due to the use of measurements at a single location and ignoring horizontal advection are smaller than 10-20% of the surface flux. The relative difference in the surface flux estimates with and without consideration of the interfacial layer depth is smaller than 5%. The results also show that the
entrainment flux is more than 50% of the surface flux in magnitude, suggesting that accurately measuring the variables at the top of CBL is important to improving the accuracy of the surface flux estimate. This is a major weakness of applying the CBL budget method because it is usually difficult to measure or predict these variables around the CBL top with ground-based instruments. In addition, the time rate of change of the mean CO$_2$ mixing ratio in the CBL is small compared to that in the NBL, and is difficult to measure accurately. Nevertheless, the method can give reasonable estimates of the diurnal and monthly variations in the surface flux. Although changes in nighttime CO$_2$ mixing ratios in the NBL are more easily detected, the large vertical gradient of the mixing ratio requires measuring vertical profiles through the NBL at high resolution. This is impractical in most of sites, and therefore it is a primary weakness of applying the NBL budget method. Under very stable conditions, the error in the estimated surface flux can be as large as 35% of the surface flux according to the comparisons between towers, while the error is only 15% under the well-mixed conditions. About 40% of data are discarded because systematic errors and non-representative sampling can occur when vertical profiles are used to infer the surface flux. As a result, only seasonal averaged NEE estimates are given, which are comparable to the tower measurements.

The vertical profiles of CO$_2$ mixing ratio measured at the tall tower enable a more accurate budget calculation both in the CBL, where the CO$_2$ mixing ratio can be directly measured in the mixed layer, and in the NBL, where the vertical profiles through the NBL can be measured in many cases. The reasonable estimates of the regional NEE suggest that it is technically promising to apply the CO$_2$ mixing ratio profiles measured at
the tall tower to inferring long-term regional fluxes. More accurate evaluation could be made if the measurements were expanded in the following ways:

(1) The CO₂ mixing ratio just above the CBL needs to be measured; or models need to be developed to predict it as a function of variables routinely measured.

(2) Approaches to estimate the mean vertical velocity at the CBL top with more confidence are needed.

(3) Measurements of CO₂ mixing ratio within the CBL need to be made at more locations to consider the effects of horizontal heterogeneity on regional scales.

(4) With respect to nighttime NEE, more efforts are needed. For example, the vertical profiles can be measured at more than one location to quantify the effects of the advection and to estimate regional NEE with more confidence.
Table 4.1: Parameters and their standard deviations for the best model among Eq. (4.11) through Eq. (4.13) describing the responses of NEE measured at the three towers and inferred regional NEE in the day to PAR or air temperature.

<table>
<thead>
<tr>
<th>Towers</th>
<th>Para</th>
<th>Growing Season</th>
<th>Dormant Season</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reg.</td>
<td>α</td>
<td>-0.005±0.0006</td>
<td>-0.030±0.0068</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-32.4±4.07</td>
<td>-34.2±7.06</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>1.22±0.36</td>
<td>3.63±0.81</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.74</td>
<td>0.78</td>
</tr>
<tr>
<td>WLEF396</td>
<td>α</td>
<td>-0.026±0.016</td>
<td>-0.03±0.009</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-13.3±2.23</td>
<td>-26.8±3.53</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>2.15±1.64</td>
<td>3.58±1.22</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.91</td>
<td>0.77</td>
</tr>
<tr>
<td>WLEF122</td>
<td>α</td>
<td>-0.052±0.022</td>
<td>-0.069±0.015</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-8.13±0.77</td>
<td>-18.6±0.88</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>3.13±0.79</td>
<td>5.67±0.92</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.71</td>
<td>0.78</td>
</tr>
<tr>
<td>WLEF30</td>
<td>α</td>
<td>-0.016±0.004</td>
<td>-0.04±0.0064</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-8.06±0.52</td>
<td>-16.2±0.61</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>2.11±0.34</td>
<td>4.51±0.54</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.90</td>
<td>0.82</td>
</tr>
<tr>
<td>LC</td>
<td>α</td>
<td>-0.016±0.013</td>
<td>-0.032±0.0072</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-1.62±0.31</td>
<td>-11.5±0.57</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>1.49±0.32</td>
<td>2.44±0.56</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.34</td>
<td>0.86</td>
</tr>
<tr>
<td>WC</td>
<td>α</td>
<td>-0.022±0.01</td>
<td>-0.072±0.01</td>
</tr>
<tr>
<td></td>
<td>P_m</td>
<td>-8.67±0.88</td>
<td>-28.2±8.88</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>2.84±0.76</td>
<td>5.01±0.91</td>
</tr>
<tr>
<td></td>
<td>r²</td>
<td>0.86</td>
<td>0.83</td>
</tr>
</tbody>
</table>

Notes: shaded cells represent the model of NEE=ae^{bT} is used, where T is air temperature at 30m in unit of ºC. a is in unit of µmol C/m²/s; b:1/ºC; NEE, R, and P_m are in unit of µmol C/m²/s; α: µmol C/µmol quanta. #: the linear model is used, NEE=α×PAR+R, where PAR is in unit of µmol/m²/s. Data from winter months are mostly in Mar, Nov, and Dec. There are not many available data in Jan and Feb. N/A: either no data, or too small r² (<0.2).
Table 4.2: Estimated parameters and their standard deviations for the model fitted to nighttime NEE in Eq. (4.13)

<table>
<thead>
<tr>
<th>Seasons</th>
<th>Methods</th>
<th>a (µmol m⁻¹ s⁻²)</th>
<th>b (°C⁻¹)</th>
<th>( r^2 )</th>
<th>( R_{10} ) (µmol m⁻¹ s⁻²)</th>
<th>( Q_{10} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dormant Season (Oct-Apr)</td>
<td>ABL budget</td>
<td>0.76±0.18</td>
<td>0.078±0.021</td>
<td>0.48</td>
<td>1.66</td>
<td>2.187</td>
</tr>
<tr>
<td></td>
<td>WLEF</td>
<td>0.93±0.04</td>
<td>0.075±0.005</td>
<td>0.90</td>
<td>1.97</td>
<td>2.117</td>
</tr>
<tr>
<td></td>
<td>Willow Creek</td>
<td>1.07±0.02</td>
<td>0.074±0.002</td>
<td>0.85</td>
<td>2.24</td>
<td>2.094</td>
</tr>
<tr>
<td></td>
<td>Lost Creek</td>
<td>0.40±0.01</td>
<td>0.078±0.003</td>
<td>0.36</td>
<td>0.87</td>
<td>2.184</td>
</tr>
<tr>
<td>Growing Season (May-Sep)</td>
<td>ABL budget</td>
<td>1.44±0.28</td>
<td>0.070±0.012</td>
<td>0.79</td>
<td>2.91</td>
<td>2.02</td>
</tr>
<tr>
<td></td>
<td>WLEF</td>
<td>2.04±0.08</td>
<td>0.062±0.002</td>
<td>0.96</td>
<td>3.79</td>
<td>1.86</td>
</tr>
<tr>
<td></td>
<td>Willow Creek</td>
<td>1.57±0.06</td>
<td>0.047±0.004</td>
<td>0.78</td>
<td>2.52</td>
<td>1.61</td>
</tr>
<tr>
<td></td>
<td>Lost Creek</td>
<td>1.16±0.07</td>
<td>0.063±0.003</td>
<td>0.54</td>
<td>2.18</td>
<td>1.87</td>
</tr>
</tbody>
</table>

Notes: The temperature at the LC tower is measured at 10m above the ground; Others are measured at 30m above the ground; At the LC tower, the respiration rate is significantly related to the depth of the water table, which is not reflected in the model used. The value of \( r^2 \) is the coefficient of determination, indicating the degree to which the changes in bin-averaged NEE with temperature can be explained by the model. \( R_{10} \) is the respiration rate at 10°C and \( Q_{10} \) is the fractional change of respiration rate in response to temperature change in 10°C.
Figure 4.1: A schematic diagram of the structure of the well developed CBL in terms of the typical vertical profile of CO₂ mixing ratio in the day under fair weather conditions during the growing season.
Figure 4.2: Typical vertical CO$_2$ mixing ratio profiles from near the surface through the lower troposphere at the WLEF site. Symbols are measurements from the powered parachute (Schulz et al., 2004), showing the profiles as a function of time in a day of summer. Lines are from aircraft measurements by NOAA/CMDL, showing typical distribution of CO$_2$ mixing ratio with height in spring, summer, and autumn. LST stands for local standard time.
Figure 4.3: Correlation coefficient between the CO$_2$ mixing ratio and the measurement height above the CBL against the standard deviation of CO$_2$ mixing ratio in the air layer between the CBL top and the height of about 3500m. The dashed line stands for the standard deviation of 0.5 ppm that is close to the precision of the measurements.
Figure 4.4: Diurnal average CO₂ jump ($\Delta C = [c]_t - [c]_m$), vertical advection term ($w_\Delta C$), and entrainment flux due to changes in CBL top ($-\frac{\partial C}{\partial t}$) in the day of each month. Winter months include November through March. Not many data are available in January and February. The CO₂ jump (solid line) is scaled by the left axis, while others are scaled by the right axis. The error bars are the standard deviations of the means.
Figure 4.5: Diurnal average of the calculated surface CO$_2$ flux, entrainment flux, and the time change rate term in each month. The diurnal average of the surface flux measured at 396m of the WLEF tower using the eddy-covariance method is shown as circles for comparison. No tower data are available in April. Winter months (a) include November through March, but not many data are available in January and February. The Error bars are the standard deviations of the means.
Figure 4.6: Comparison of the monthly diurnally-averaged regional NEE inferred from the CBL budget method and NEE measured at the three levels of the WLEF tower using the eddy-covariance method. The variables $y_{30}$, $y_{122}$, and $y_{396}$ in the linear equations represent NEE measured at 30, 122, and 396m of the WLEF tower, respectively, while $x$ represents NEE from the CBL budget method.

$y_{30} = 0.40x - 1.60$
$y_{122} = 0.58x - 0.86$
$y_{396} = 0.71x - 0.87$
Figure 4.7: Bin-averaged NEE inferred from the CBL budget method and directly measured at the WC, LC, and WLEF towers as a function of PAR in each month, (a) November through March, in this case, not many data are available in Jan and Feb; (b) April, (c) May, (d) June, (e) July, (f) August, (g) September, (h) October. Data are binned by PAR in an increment of 200 µmol m$^{-2}$s$^{-1}$. The standard deviation is for the mean NEE in each bin. The values of $r^2$ are the coefficients of determination, indicating the degree to which the changes in bin-averaged NEE with PAR can be explained by the fitted curves (thick lines). Parameters of the light response curve are reported at Table 4.1.
Figure 4.8: Bin-averaged NEE derived from the CBL budget method and directly measured at the three towers varying with air temperature at 30m (except the LC tower where air temperature is measured at 10m) during the dormant season from October to April. Data are binned by air temperature in an increment of 4°C. Note that there are not many data available in January and February. Parameters of the model are reported in Table 4.1. The Error bars are the standard deviations of the means.
Figure 4.9: Comparison of monthly-averaged daytime NEE derived from the CBL budget method, and measured at three levels of the WLEF tower, Willow Creek tower, and Lost Creek tower during the growing season in 2000 and 2003. Data are available only in 2003 at the Lost Creek tower. ‘NG’ stands for non-growing season, where NEE values are averaged in the months from October to April. The error bars are the standard deviations of the means.
Figure 4.10: (a) The median value of binned nighttime NEE calculated from the one-dimensional budget method with tall tower measurements by $u_*$ in increment of 0.025 m s$^{-1}$ as a function of the friction velocity. Error bars are standard deviations of the binned data. The bin averages and standard deviations of soil temperature at 5 cm below the surface and air temperature at 30 m above the surface are also presented by triangles and filled circles, respectively. (b) Normalized nighttime NEE as a function of $u_*$ and frequency of data in each bin of $u_*$. 
Figure 4.11: (a) Nighttime NEE, soil temperature, and air temperature as a function of Monin-Obukov length $L$. All data are binned by the logarithm to the base 10 of $L$ in increment of 0.2. (b) Normalized NEE as a function of Monin-Obukov length and distribution of frequency of the data in each bin. The error bars are the standard deviations of the means.
Figure 4.12: Vertical turbulent fluxes of CO₂ measured at 30m, 122, and 396m during the growing season against (a) the friction velocity scale, $u^*$, and (b) Monin-Obukov length $L$. The error bars are the standard deviations of the means.
Figure 4.13: The relative difference (RD) of the time change rate terms below 30m at the WLEF and WC sites under low (squares) and high (filled circles) $u^*$ conditions. RD is calculated as the difference in the integral of the time derivative of CO$_2$ mixing ratio from the surface to 30m at the two towers divided by the NEE measured at the WLEF tower. The monthly mean NEE measured at the WLEF tower, scaled by the right axis, is also shown. The error bars are the standard deviations of the means.
Figure 4.14: Nighttime NEE estimated from the budget method and from tower measurements as a function of air temperature at 30m above the ground during the dormant season (a) and growing season (b). All data are binned by temperature in an increment of 2°C. The error bars are the standard deviations of the means.
Figure 4.15: Deviation of the time-derivative of CO$_2$ mixing ratio measured at 76m above the ground at the WLEF tower from the mean time derivative measured at the same height at the WLEF and other two tall towers in the region (see text) as a function of $u^*$ under fair weather conditions in August 2003. Squares and filled circles are for the cases in the day and at night, respectively.
Figure 4.16: Comparison of seasonally-averaged nighttime NEE derived from the ABL budget method, measured at the WLEF, WC, and LC towers, respectively. The error bars are the standard deviations of the means.
Chapter 5

Assessing Effects of Horizontal Advection on Regional CO₂ Flux Estimates

5.1 Introduction

The convective boundary layer (CBL) budget method is usually employed to estimate the surface fluxes of carbon dioxide (CO₂) on the spatial and temporal scales of $10^2$ km and one hour, respectively (Batchvarona et al., 2001; Denmead et al., 1996; Gryning and Batchvarova, 1999; Kuck et al., 2000; Lloyd et al., 2001; Raupach et al., 1992; Soegaard, 1999). Of all the terms in the conservation equation of CO₂ mixing ratio, the effects of the horizontal advection terms, however, are usually ignored and not well quantitatively assessed in the literature. Data from the current network of the measurements of the mixing ratio and flux of CO₂ cannot be used directly to assess such effects at least for two reasons. First, the network is too sparse. Second, the instruments are not well calibrated at most sites. In fact, it is not readily to measure the horizontal gradients of CO₂ mixing ratio in the field mainly due to the requirements of the high calibration accuracy and precision for instruments. The dependence of the effects of the horizontal advection on spatial scales also complicates experimental designs for quantitatively assessing them. As a result, the horizontal gradients are usually not directly
measured in practice except in special campaigns (e.g., http://rflux.psu.edu). In this case, one would turn to other approaches to achieve the assessment.

Observational studies (Baldocchi, 1994; Egan et al., 1991) indicated that the mixing ratios and surface fluxes of water vapor and CO$_2$ are usually coupled over vegetated areas due to the physiological activities of plants. According to those studies, one may hypothesize that the measurements of the mixing ratio and surface flux of water vapor could provide a constraint for estimating both the surface flux and the distribution of the mixing ratio of CO$_2$ in the atmosphere. With the hypothesis under certain conditions, we can take advantage of the observational networks for water vapor mixing ratio, which are much denser than those for CO$_2$ mixing ratio in terms of the number of the observational stations. For example, water vapor mixing ratio data can be easily acquired from routine meteorological observation networks such as routine surface and upper air meteorological networks (McIlveen, 1992) and regional environmental monitoring networks (e.g., Brock et al., 1995). Efforts are made to examine such a hypothesis in this study.

There are two objectives of this study. One is to assess the effects of the horizontal advection on CO$_2$ surface flux estimates using aircraft data on the temporal and spatial scales, respectively, of one hour and 20km that are in the range where the CBL budget method often applies. The other is to explore the possibility to evaluate such effects indirectly in terms of water vapor mixing ratio data. Section 5.2 describes the data used and compares the contributions of the horizontal advection terms with the surface fluxes for CO$_2$ and water vapor, respectively. Starting with a hypothesis implied from data analyses, we identify a scenario in section 5.3, under which the horizontal gradients
of CO₂ mixing ratio can be constrained by the measurements for water vapor mixing ratio in some cases. Section 5.4 presents an application of the constraint to evaluate the effects of the horizontal advection on regional CO₂ flux estimates over a forested area in northern Wisconsin. Finally, discussions and summary are presented.

5.2 Evaluating the effects of the horizontal advection on surface flux estimates using aircraft data

5.2.1 Budget equation

The surface flux, denoted by \( F_c \), for a trace gas such as CO₂ or water vapor can be written in the mean wind direction coordinate as, by integrating the conservation equation for the mixing ratio of the gas from the surface to the top of the CBL,

\[
F_c = F_{\text{top}} + \int_0^h \frac{\partial [c]}{\partial t} \, dz + h \left[ [u] \frac{\partial [c]}{\partial x} \right]_m + \int_0^h [w] \frac{\partial [c]}{\partial z} \, dz , \tag{5.1}
\]

where \( F_{\text{top}} \) is the vertical flux at the top of the CBL; \( h \) is the CBL depth, \([c]\), \([u]\), and \([w]\) represent the mean mixing ratio of the gas, mean horizontal and vertical wind speeds, respectively. The third term on the right-hand side of the equation is called the horizontal advective flux term hereafter, representing the contribution of the horizontal advection to the estimate of the surface flux, which is examined for CO₂ and water vapor below.
5.2.2 Description of the experimental site and data

The data recorded at The University of Wyoming King Air (UWKA) aircraft during the IHOP experiment is used to evaluate the effects of the horizontal advection. The experiment took place over the Southern Great Plains of the continental U.S. from middle May to middle June in 2002. Figure 5.1 presents the locations of surface stations and three flight tracks, i.e., west track, central track, and east track (solid lines). The west track consists of fallow, grassland, and sagebrush; the central track consists mostly of winter wheat in addition to grassland; the east track consists of grassland. The UWKA aircraft flew at about the top of the surface layer (70-80m), the middle of the ABL (300-700m), and near the top of ABL (800-1500m) over the three tracks. Each flight track is about 60-80km long. The flight period for each day is mostly between the late morning and the middle afternoon. During the 14 flight days, the winds contained significant components parallel to the tracks only at some hours on 4 days, allowing the examination of the impacts of the horizontal advection.

The mean mixing ratios and fluxes of CO$_2$ and water vapor are computed from aircraft data for a window of 20km using unweighted mean. The window is sequentially marched through the record by an increment of 1km. The mean values of NDVI (Normalized Difference of Vegetation Index) for the west, central, and east tracks are about 0.1, 0.4, and 0.65, respectively. The CBL depths are estimated from aircraft soundings. Details for the site information and IHOP can be found in the website of http://www.atd.ucar.edu/dir_off/projects/2002/IHOP.html.
5.2.3 Observed effects of the horizontal advection

The flights on May 29 and June 16 2002 are selected as case studies. The wind direction is thought of as being parallel to the flight track under the condition of $|\sin \Phi| < 0.2$, where $\Phi$ is the angle between the flight and wind directions. In the two days, approximately half of the flights were made at the top of the surface layer, allowing us to calculate both the horizontal advection components and the surface fluxes. In the calculation, the horizontal advection term (i.e., $\frac{\partial \theta}{\partial x}$) is assumed to be approximately uniform with height in the CBL. As a result, the horizontal advective flux term in the budget equation is equal to the product of the CBL depth and the measured horizontal advection of the gas. Figure 5.2 presents the horizontal advective fluxes and surface fluxes for CO$_2$ and water vapor calculated on the spatial scale of 20km.

On May 29, five of nine flights over the west track parallel to the wind directions were made at the top of the surface layer, hence providing the estimates of the surface fluxes. The other four flights are made at the middle of the CBL. The magnitudes of the horizontal advective fluxes of CO$_2$ are about 0.1 ppm m/s, four times larger than those of the surface fluxes of about 0.02 ppm m/s (Figure 5.2a). The small magnitudes of the surface fluxes indicate that the effects of horizontal advection are compensated mostly by other components, implying that ignoring them would lead to significant errors (>100%) in the estimates of the surface fluxes using the budget equation. With respect to water vapor, the mean surface flux is about 0.1 g/kg m/s, while the horizontal advection term is about -0.1 g/kg m/s, of the same order of the surface flux. The significant effects of the
horizontal advection in this case are due probably to both the heterogeneous surface fluxes and the resulting complex structure of the boundary layer (Kang, in preparation).

In contrast, the surface condition of the central track is more homogeneous in terms of the spatial distribution of vegetation. On June 16, four flights were made over the central track parallel to the wind directions. Two of them were made at the top of the surface layer with the other two flights at the middle of the CBL. The horizontal advection term for CO$_2$ is, on average, about -0.05 ppm m/s, which is 25% of the average surface flux of about -0.2 ppm m/s (Figure 5.2b), indicating that the surface fluxes would be overestimated by 25% if the effects of the horizontal advection were ignored. With respect to water vapor, the average horizontal advective flux is about 0.03 g/kg m/s with the surface flux being about 0.1g/kg m/s. Compared with those over the west track, the effects of the horizontal advection over the central track are smaller.

5.2.4 Implication

The magnitude of the horizontal advective flux is a function of spatial scales, which can be shown readily by a scaling argument, i.e.,

$$ h \left[ u \frac{\partial [c]}{\partial x} \right] \sim H \frac{U \Delta C}{L}, \quad (5.2) $$

where $H$, $U$, and $\Delta C$ are the characteristic CBL depth, horizontal wind speed, difference in CO$_2$ mixing ratio over the horizontal spatial scale, $L$. Eq. (5.2) indicates that the effect of the horizontal advection can decrease with the spatial scale. However, with increasing spatial scale, the measurements at one location, e.g., at a tall tower or tethered balloon,
are less representative in the region. In practice, the horizontal scale on which the CBL budget method usually applies is of the order of $10^2$ km (typically about 50km) in consideration of the capability of the mixing processes in the CBL (Denmead et al., 1996; Lloyd et al., 2001; Raupach et al., 1992). The evaluation in the previous section implies that on the spatial scale of that order, the effects of the horizontal advection can still be significant on the temporal scale of one hour. Therefore, the effects of the horizontal advection need to be evaluated to estimate the regional surface fluxes with more confidence using the budget method. An approach is proposed below to provide a constraint for the effects of the horizontal advection on regional CO$_2$ flux estimates in terms of the horizontal gradients of water vapor mixing ratio when reliable and dense measurements of CO$_2$ mixing ratio are unavailable.

5.3 Constraining the effects of the horizontal advection

The goal of this section is to find a possible relationship between the horizontal gradients of water vapor and CO$_2$ mixing ratios. We begin with a hypothesis implied from data analyses. Then a scenario under which the hypothesis can apply is identified.

5.3.1 Phenomenon and hypothesis

Results in Figure 5.2 show that the signs of the horizontal advective fluxes of CO$_2$ and water vapor are opposite like those of their corresponding surface fluxes, suggesting
likely correlations between the horizontal gradients and surface fluxes. As an effort to find the relationship between the horizontal gradients of water vapor and CO₂ mixing ratios, the following equation is examined by using aircraft data,

\[
\frac{\partial \bar{c}}{\partial t} + [u] \frac{\partial \bar{c}}{\partial x} = \alpha \frac{F_c}{F_q}
\]

(5.3)

where \( \alpha \) is a coefficient; \( F_c \) and \( F_q \) are the surface fluxes for CO₂ and water vapor, respectively; \( \bar{c} \) and \( \bar{q} \) are the spatial average mixing ratios of CO₂ and water vapor, respectively. The ratio of \( F_c \) and \( F_q \) is also called water use efficiency (WUE). Figure 5.3 compares the leg-mean WUE with the ratio in the left hand side of Eq. (5.3) over three time periods between noon and the middle afternoon for the flights when the wind directions are approximately parallel to the flight tracks. In the figure, the values of \( \alpha \) can be regarded as the slopes of the lines linking the corresponding data points with the origin. For the flights over the higher NDVI area (NDVI=0.4), the coefficient, \( \alpha \), is confined in a range between 0 and 1, tending to be close to unity in the middle afternoon. In contrast, for the flights over the west track with smaller NDVI, the coefficient can be any value. At first glance, the difference in the behaviors of the coefficients might be due, in part, to the different structures of the boundary layer over the two areas (Kang, in preparation) as well as the impacts of the surface vegetation. Therefore, based on the scenarios when the experiments were conducted and the results in Figure 5.3, we can hypothesize that the coefficient can be confined in a range from 0 to 1 over a vegetated area with horizontally homogenous boundary layer under fair weather conditions in the afternoon. Note that the hypothesis is used only to interpret the cases with high NDVI, though the coefficient, \( \alpha \), may also be between 0 and 1 in some cases with low NDVI.
5.3.2 Interpretation

In this section, efforts are made to define a scenario where the above hypothesis holds. Determining when the scenario is met, we can find a constraint for the horizontal gradients of CO$_2$ mixing ratio in terms of the measurements of water vapor mixing ratio.

We focus our discussion on daytime CBL with large-scale vertical atmospheric subsidence under fair weather conditions. The situations when clouds are presented at the CBL top are not considered. The CBL structure is assumed to be horizontally homogeneous, implying that the effect of the surface heterogeneities on the bulk properties of the CBL is negligible. A typical scenario under those assumptions is that an area is controlled by a high pressure system with the bulk of the troposphere consisting of dry, slowly subsiding air. In this case, the height of CBL is usually reasonably predictable. These conditions are required to apply the CBL budget method as well.

Considering a slab model of the convective boundary layer under the above conditions, as usually employed in the literature (e.g., Denmead et al., 1996; Garratt, 1992; Lloyd et al., 2001; Raupach et al., 1992), we can rewrite Eq. (5.1) for CO$_2$ as,

$$F_c = h \frac{\partial [c]_m}{\partial t} + h \left[ u \left( \frac{\partial [c]}{\partial x} \right)_m - \left( \frac{\partial h}{\partial t} - w_+ \right) ([c(h)]_+ - [c]_m) \right)$$  \hspace{1cm} (5.4)

where $w_+$ is the mean vertical velocity at the top of the CBL, typically induced by the large scale subsidence of airflow; $[c(h)]_+$ and $[c]_m$ are the mean mixing ratios of CO$_2$ just above the CBL top and within the CBL, respectively. The last term of the right-hand side of Eq. (5.4) can be expressed as,
where $\Delta[c]$ is equal to $[c(h)]_{-}-[c]_{m}$, the difference in the mixing ratio of CO$_2$ just above and within the CBL; $k_{c}$ is the ratio of the entrainment flux to the surface flux for CO$_2$. Eq. (5.5) is an expression of the entrainment flux at the top of the CBL, characterizing the exchange of the CO$_2$ within and above the CBL (Garratt, 1992). Rearranging Eq. (5.5) yields,

$$k_{c} = \frac{- \left( \frac{\partial h}{\partial t} - w_{*} \right) \Delta[c]}{F_{c}}.$$  \hspace{1cm} (5.6)

Substituting Eq. (5.5) into Eq. (5.4) yields,

$$h \frac{\partial [c]_{m}}{\partial t} + h \left[ u \frac{\partial [c]}{\partial x} \right]_{m} = (1 - k_{c})F_{c}.$$  \hspace{1cm} (5.7)

Similar to Eq. (5.6) and Eq. (5.7), we can write the equations for the water vapor mixing ratio, i.e.,

$$k_{q} = \frac{- \left( \frac{\partial h}{\partial t} - w_{*} \right) \Delta[q]}{F_{q}},$$  \hspace{1cm} (5.8)

and,

$$h \frac{\partial [q]_{m}}{\partial t} + h \left[ u \frac{\partial [q]}{\partial x} \right]_{m} = (1 - k_{q})F_{q},$$  \hspace{1cm} (5.9)

where $\Delta[q]$ is equal to $[q(h)]_{-}-[q]_{m}$, the difference in the mixing ratio of water vapor just above and within the CBL; the variables with the subscript $q$ have the same meanings as those in Eq. (5.6) and Eq. (5.7) except for water vapor. Taking the ratio of Eq. (5.7) and Eq. (5.9) yields,
Notice that the mixing ratios used to examine Eq. (5.3) are measured at the top of
the surface layer or in the middle of the CBL, approximately representing the
 corresponding mean values in the CBL. Therefore, the term \((1-k_c)/(1-k_q)\) is equal to the
coefficient, \(\alpha\). In other words, the term \((1-k_c)/(1-k_q)\) is possibly confined between 0 and 1
according to the hypothesis, which is interpreted starting with an ideal case below.

5.3.2.1 One-dimensional case under ideal conditions

Without the effects of the advection terms being taken into account, the budget
equation for CO\(_2\) is,

\[
\frac{h \partial [c]}{\partial t} + \frac{h \partial [g]}{\partial t} + u \frac{\partial [c]}{\partial x} \left|_m \right. = \frac{(1-k_c)}{\left(1-k_q\right)} F_c .
\]  \hspace{1cm} (5.10)

Integrating Eq. (5.11) from \(t=t_{c0}\) to \(t\) yields,

\[
\int_{t_{c0}}^{t} F_c \, dt = -h(t) \Delta [c] + \Gamma_c I_c (t) ,
\]  \hspace{1cm} (5.12)

where \(I_c(t) = \left(h(t)^2 - h(t_{c0})^2\right)/2 - \int_{t_{c0}}^{t} w_c h dt\) is positive due to negative \(w_c\); \(\Gamma_c\) is the
vertical gradient of the CO\(_2\) mixing ratio above the CBL, which is usually positive in the
growing season (Figures 4.2 and 4.3 in Chapter 4); \(t_{c0}\) is the initial time of the
integration. In deriving Eq. (5.12), three assumptions have been made (see Appendix C.1
for details). First, the effects of the horizontal advection of CO₂ above the CBL are negligible compared to those of the vertical advection. Second, the CO₂ mixing ratio is distributed nearly linearly with height above the CBL. Third, the initial time $t_{c0}$ is chosen as the time when $[c]_m$ is equal to $[c]_+$, usually between the early and middle morning when plants assimilate the respired CO₂ accumulated in the nocturnal boundary layer. As an example, Figure 5.4a presents the vertical profiles of the CO₂ mixing ratio evolving with time measured over a forested area in a day with fair weather, indicating the assumptions are possibly reasonable in practice.

With respect to water vapor, similar assumptions can be made. The resulting time-integrated water vapor flux is,

$$\int_{t_{c0}}^t F_q dt = -h(t)\Delta[q] + \Gamma_q I_q,$$  \hspace{1cm} (5.13)

where $I_q = \left( h(t)^2 - h(t_{q0})^2 \right) / 2 - \int_{t_{q0}}^t w_t h dt$ is positive like $I_c$; $\Gamma_q$ is the vertical gradient of the water vapor mixing ratio above the CBL; $t_{q0}$ is the initial time of the integration when $[q]_m=[q(h)]_+$ or $h=0$. It should be noted that $t_{q0}$ is not necessarily equal to $t_{c0}$. The impact of the difference in the two initial times on the results is discussed later. As discussed in Raupach et al. (1992), the water vapor mixing ratio above the CBL is well constrained in the dry, subsidence weather patterns. As a result, there is no significant change in the water vapor mixing ratio with height above the CBL, i.e., $\Gamma_q$ is close to zero (Figure 5.4b).

In the one-dimensional case, $k_c$ and $k_q$ can be written as,
According to Eq. (5.14) and Eq. (5.15), the ratio of $k_c$ and $k_q$, denoted by $r(t)$, is,

$$r(t) = \frac{k_c}{k_q} = \frac{\Delta[c]}{\Delta[q]}$$

With Eq. (5.12) and Eq. (5.13), $r(t)$ can be further written as,

$$r(t) = \frac{k_c}{k_q} = \frac{\Delta[c]}{\Delta[q]} \frac{F_q}{F_c}$$

where the inequality is due to the opposite signs of $F_c$ and $\Gamma_c$ observed over vegetated areas typically in the growing season and small $\Gamma_q$ (see Appendix C.1 for the detailed evaluation). $WUE(t)$ is the water use efficiency at time $t$, i.e.,

$$WUE(t) = \frac{F_c(t)}{F_q(t)}$$

The value of the term on the far right hand side of Eq. (5.17) depends largely on the behaviors of the evolutions of $F_q(t)$ and $WUE(t)$ with time. Under fair weather conditions, $F_q(t)$ usually increases with time till the middle afternoon without the change in sign. In contrast, the behavior of the variation of $WUE$ varies with seasons and surface conditions. In the growing season, the magnitude of $WUE$ generally decreases as the
vapor pressure deficit (VPD) increases over vegetated areas (e.g., Baldocchi, 1994; Mahrt and Vickers, 2002). As VPD increases, the partial stomatal closure physiologically reduces the CO₂ flux more than the water flux. Because VPD usually increases from morning to mid-afternoon in the atmosphere under fair weather conditions, the magnitude of WUE is observed typically to be decreasing with time from the morning to mid-afternoon. As an example, Figure 5.5 presents the monthly mean WUE varying with time at a forested site in each month during the growing season. By contrast, the patterns of the WUE variation are not well defined during the dormant season.

With the variation behaviors of WUE and \( F_q \), it can be readily shown that \( r \) is greater than unity during the period from the morning to middle afternoon during the growing season over vegetated areas in the case of \( t_{c_0} \leq t_{q_0} \), i.e.,

\[
k_c(t) > k_q(t).
\]

(5.19)

In the case of \( t_{c_0} > t_{q_0} \), \( r(t) \) can still be greater than unity if \( t \) is far later than \( t_{q_0} \), e.g., in the late morning or early afternoon, as a result of the small magnitudes of \( F_q \) in the early morning compared with those in the middle day (see appendix C.1). In practice, the initial times \( t_{c_0} \) and \( t_{q_0} \) are difficult to quantitatively define due to variable biological and atmospheric conditions case by case, though they can appear in the early morning, e.g., between 0600 and 0800 local time. Consequently, to ensure that Eq. (5.19) holds regardless of the initial integration times, time \( t \) should be selected as late as possible during the period when WUE decreases in magnitude, e.g., early afternoon.

In addition, both \( k_c \) and \( k_q \) are greater than or equal to zero during the CBL developing period according to their definitions for two reasons. First, the time rate of
change in the depth of the CBL is greater than zero during the period (Figure 5.6). As the CBL depth reaches the maximal value, time derivative of \( h \) is equal to zero, resulting in zero \( k_c \) and \( k_q \). Second, the surface fluxes \( F_c \) and \( F_q \) usually have opposite signs to \( \Delta[c] \) and \( \Delta[q] \), respectively. Therefore, Eq. (5.19) can be rewritten as,

\[
k_c(t) \geq k_q(t) \geq 0,
\]

where the equalities hold at the CBL depth reaches its maximal value. Eq. (5.20) provides a lower bound for \( k_c \), which can hold as well even with the advection terms being considered in some cases.

### 5.3.2.2 Considering the advection terms

Considering the advection terms, we can rewrite Eq. (5.12) and Eq. (5.13) as,

\[
\int_{t_a}^{t} F_c' dt = -h(t) \Delta[c] + \Gamma_c I_c,
\]

and,

\[
\int_{t_a}^{t} F_q' dt = -h(t) \Delta[q] + \Gamma_q I_q,
\]

where \( F_c' = F_c - \left[ u \frac{\partial[c]}{\partial x} \right]_m - w_c \Delta[c] \) and \( F_q' = F_q - \left[ u \frac{\partial[q]}{\partial x} \right]_m - w_q \Delta[q] \).

The ratio of \( k_c \) and \( k_q \) then can be written as,
The inequality \( r(t) > 1 \) can still hold in the afternoon for the cases with the moderate effects of advection over vegetated surfaces, depending on the time rate of the change in \( WUE \) as shown in Appendix C.2. For example, it is possible that \( r(t) \) is greater than unity over a forested area (the WLEF area) in the afternoon if the magnitude of the overall total advective flux in the horizontal and vertical directions is smaller than 40% to 80% in comparison with that of the surface flux.

Although Eq. (5.20) and Eq. (5.23) provide an estimated lower bound for \( k_c \), the water vapor measurements alone cannot define an upper bound for \( k_c \). A crude way is to find conditions when \( k_c \) is smaller than an imposed upper bound with a scaling argument. With Eq. (5.21), \( k_c \) can be approximated as (see Appendix C.3 for derivation),

\[
k_c(t) \approx \frac{\left( \frac{\partial h}{\partial t} - w_+ \right) \Delta t}{h(t)} \leq k,
\]

where \( k \) is the imposed upper bound; \( \Delta t \) is the difference between \( t \) and \( t_{c_0} \).

Given the CBL depth, its time derivative, and \( k \), we can find a time period when the inequality Eq. (5.24) is met. For example, with typical values in the middle day, \( h=1500\text{m}, \Delta t=6\text{ hours}, w_+=-0.01\text{m/s}, \) and \( k=1 \), we have,

\[
\frac{\partial h}{\partial t} \leq \frac{hk}{\Delta t} + w_+ = 214 \text{ m/hour},
\]

(5.25)
which can be found during the period from noon to the afternoon because \( \frac{dh}{dt} \) usually decreases with time before the CBL depth reaches its maximal value (Figure 5.6). Therefore, the time period when \( k_c \) is smaller than \( k \) can be found given the measurements or predictions of the CBL depth in practice. With \( k_q < k_c < 1 \), we have,

\[
0 < \frac{1 - k_c}{1 - k_q} < 1,
\]

which is indicated in Figure 5.3. It should be noted that the information on the upper bound is limited, as the scaling argument can only give an estimate of the order of \( k_c \). For more definable estimates, other constraints are needed. For example, we can use another type of trace gas whose features can constrain \( k_c \) with an upper bound in the similar way like Eq. (5.20).

### 5.4 Application

The effects of the horizontal advection on the estimates of CO\(_2\) surface fluxes using the CBL budget method are assessed in a forested region in Northern Wisconsin based on the hypothesized constraint in this study.

#### 5.4.1 Site and data

The 40×40 km\(^2\) region covered by a mixed forest is centered at the WLEF tall tower (Berger et al., 2001). The dominant forest types are mixed forest uplands and wetlands. The eddy-covariance fluxes of sensible heat, latent heat, and CO\(_2\) are measured
at three levels of the tower. CO$_2$ mixing ratios are measured at six levels, i.e., 11, 30, 76, 122, 244, and 396m above the ground (Bakwin et al., 1998), which are used to infer surface fluxes with the CBL budget method (Chapter 4). The spatial distributions of the mixing ratios of water vapor are obtained from the RUC (rapid updated cycle) reanalysis data (Benjamin et al., 2004). The terrain in the region is relatively flat but the distribution of surface vegetation is heterogeneous. The horizontal advective fluxes during the growing seasons in 2000 and 2003 are evaluated. The CBL height is estimated by using an empirical model developed with data from a 915MHz radar (Yi et al., 2001).

5.4.2 Case study

Given the measurements of the mixing ratio of water vapor, estimated $WUE$, and the CBL depth, a lower bound for the effects of the horizontal advection, denoted by $F_{\text{Lower}}$, in the afternoon can be estimated as, by rearranging Eq. (5.10) with $k_c = k_q$,

$$F_{\text{Lower}} = WUE \times \left\{ h \frac{\partial[q]}{\partial t} + h \left[ u \frac{\partial[q]}{\partial x} \right]_m \right\} - h \frac{\partial[c]}{\partial t} \leq h \left[ u \frac{\partial[c]}{\partial x} \right]_m \equiv F_{\text{hor}}, \quad (5.27)$$

where $F_{\text{hor}}$ represents the horizontal advective flux. With $k_c = 1$, we have an upper bound for the horizontal advective flux, $F_{\text{Upper}}$, i.e.,

$$F_{\text{Upper}} = -h \frac{\partial[c]}{\partial t} > h \left[ u \frac{\partial[c]}{\partial x} \right]_m = F_{\text{hor}}. \quad (5.28)$$
Besides the method of using the scaling argument, an alternative upper bound for $F_{\text{hor}}$ can be estimated provided that the values of $F_{\text{hor}}$ have nearly the same magnitudes but reversed signs in the opposite directions, i.e.,

$$F_{\text{hor}}(\Theta) \approx -F_{\text{hor}}(\Theta + 180^\circ) \leq -F_{\text{Lower}}(\Theta + 180^\circ), \quad (5.29)$$

and thus,

$$F_{\text{Upper}}(\Theta) \approx -F_{\text{Lower}}(\Theta + 180^\circ), \quad (5.30)$$

where $\Theta$ is the wind direction in degree. Similarly, another lower bound estimated from the upper bound, Eq. (5.28), can be expressed as,

$$F_{\text{Lower}}(\Theta) \approx -F_{\text{Upper}}(\Theta + 180^\circ), \quad (5.31)$$

The assumption used in deriving Eq. (5.29) and Eq. (5.31) implies that the averaged horizontal advective fluxes can be mostly cancelled out in the opposite wind directions, which is possibly reasonable if the horizontal advection is due mainly to the heterogeneous distribution of surface vegetation on local and regional scales. There are three extreme conditions under which the assumption is invalid. First, the location of a site is close to extremely large sources or sinks, e.g., strong point sources or sinks, compared to the typical strength of surface sources or sinks in the region. In other words, if surface sources or sinks gradually vary in space, the assumption is likely to be reasonable. Second, a site is located at the center in terms of the spatial distribution of the strength of surface sources or sinks. In this case, the assumption of the reversed signs in the opposite directions is inappropriate. The above two conditions can be examined roughly by the spatial distribution of water vapor mixing ratio due to the close correlation.
between the two scalars, indicating that no such extreme conditions possibly exist in the WLEF region. Third, the effects of the horizontal advection at synoptic scales are much larger than those at local and regional scales. This phenomenon is unlikely under the atmospheric conditions we are considering (unless the horizontal advective effects on regional scales are zero). Therefore, the assumption can provide another possible constraint for the horizontal advective fluxes.

The lower and upper bounds for the horizontal advective fluxes around the WLEF tower, averaged over the growing season, are estimated (Figure 5.7) according to the water vapor constraint hypothesis in eight wind direction sectors, i.e., north (N), northeast (NE), east (E), southeast (SE), south (S), southwest (SW), west (W), and northwest (NW) directions. In the calculation, the $WUE$ and CO$_2$ mixing ratio are estimated from the tower measurements, while the horizontal gradients of the water vapor mixing ratio are estimated from the RUC reanalysis data. To ensure that the inequality Eq. (5.20) is met and that $k_c$ is smaller than unity, we select data that are between noon and 1500 local time and satisfy Eq. (5.24) with the imposed $k$ being equal to 1.

When the wind comes from NE, E, and SE directions, the average values of the low bounds (filled circles in Figure 5.7) are likely close to zero or positive, indicating that the horizontal advective fluxes are most likely positive. In other words, ignoring the contribution of the horizontal advection could underestimate the surface fluxes mathematically in those wind directions. In the N wind direction, the sign of the horizontal advective flux cannot be determined in that the probability of the lower bound being zero or positive is small (<50%). The average values of the lower bounds are negative when the wind blows from S, SW, W, and NW directions. The horizontal
advective flux is most likely negative when the wind comes from NW because the upper bound is mostly likely negative or close to zero, suggesting that ignoring the effect of the horizontal advection could overestimate the surface flux mathematically in that wind direction. In the S, SW, W wind directions, the signs cannot be determined, for the upper bounds based on the scaling argument (open squares in the figure) provide insufficient constraint information.

With the assumption that the horizontal advective fluxes have nearly the same magnitudes and reversed signs in the opposite directions, another set of the upper and lower bound values are obtained (dashed and dot dash lines in Figure 5.7, respectively), giving more constraints to the horizontal advective fluxes. Combination of the two low bounds and two upper bounds in each wind direction creates a narrower range of the horizontal advective flux, which is between the minimum of the two upper bounds and the maximum of the two lower bounds (shaded area in Figure 5.7). In this case, the horizontal advective fluxes in the SW, W, and NW wind directions are likely negative, ranging approximately from about -3 to 0 µmol m⁻²s⁻¹ in contrast with those in NE, E, and SE wind directions ranging from 0 to 3 µmol m⁻²s⁻¹. When the winds are from N and S directions, the fluxes are about between -2 and 2 µmol m⁻²s⁻¹, while the signs cannot be estimated. On average, the magnitude of the contribution of the horizontal advection is of the order of that of the vertical advection, 15-20% of the magnitude of the mean surface flux on the spatial scale of 40km and temporal scale of one hour. Most likely, the surface fluxes derived from one-dimensional budget equation are overestimated (more negative) in the NE, E, and SE wind directions, while underestimated (less negative) in the SW, W,
and NW wind directions. The range of the estimated magnitude can be narrower with more constraints.

The advantage of the method is that the contribution of the horizontal advection terms can be evaluated by measurements of the mixing ratio of CO$_2$ at a single location, and spatial distribution of water vapor mixing ratio, while the disadvantage is that the WUE measured at the tower is used, leading to uncertainty.

5.5 Discussion

Using the water vapor constraint can estimate a lower bound for the effects of the horizontal advection in the scenario identified in this study. In practice, the approach applies over vegetated areas during the growing season under most fair weather conditions, in particular in a high pressure area, where the CBL is well developed. In principle, the measurements of the mixing ratios of other tracers can also provide constraints in a similar way, but the advantage of the use of the water vapor measurements is that its measurement network is the densest and easily accessed. There are some limitations about the use of the water vapor constraint.

Firstly, the constraint holds only in the late morning or early afternoon during the growing season. It is difficult to exactly define a time threshold on a day in practice. Empirically, we can select times as late as possible based on the analyses from the scaling argument to facilitate applying the water vapor constraint. As a result of this restriction, the effects of the horizontal advection other than those times cannot be estimated. Nevertheless, the approach still can estimate a range for their overall magnitude because
the effects of horizontal advection are usually more significant in the afternoon than at other times in the day. In addition, the approach makes use of an important feature of \textit{WUE} over vegetated areas during the growing season, i.e., the magnitude of \textit{WUE} decreases with time from early morning to afternoon under fair weather conditions. The evolution pattern of \textit{WUE}, however, is not well defined in the dormant season. Therefore, it is difficult to determine scenarios to apply similar constraints during that season.

Secondly, the approach can be used when the effects of the overall total advection are moderate, e.g., smaller than 40% to 80% of the magnitude of the surface flux in the forested area shown in this study, depending in part on the evolution rate of \textit{WUE} with time. For the cases with larger effects of the advection, the approach may not work. In practice, to a first order approximation, we can compare the contribution of the advection of water vapor to its surface flux and roughly evaluate whether the effects of the advection of CO$_2$ are possibly larger than the required upper limits or not, due to their close correlation.

Thirdly, the bound values estimated from the water vapor constraint may vary with time even for similar effects of the horizontal advection in the same day. The estimated range of the effects of the horizontal advection at a specific time depends on how close the $k_q$ value is to 1. In general, the range becomes narrower with increasing $k_q$. Provided that the effects of the horizontal advection are the same in a time period, the best estimates for the lower and upper bounds would be the maximum and minimum values among the results calculated at different times, respectively. But due to unidentified random errors, it is unlikely to find them in practice. As a compromise, we take the central values as the approximations as shown in Figure 5.7, which actually
underestimate and overestimate the lower and upper bounds, respectively. In other words, the ranges for the effects of the horizontal advection shown in the figure are somewhat overestimated. The error bars for the estimated bounds in Figure 5.7 represent the total contribution of both the random errors and the variations of the results with time during the calculated period.

Fourthly, using the scaling argument to determine the upper bounds provides insufficient information. An alternative way, for instance, is to use another type of tracer that can provide upper bounds in the way similar to or different from that used in this study.

Fifthly, the assumption that the effects of the horizontal advection in opposite wind directions can be mostly cancelled out might result in the bias estimates if there are strong dependences of atmospheric variables such as the wind speed and CBL depth on wind directions, or if the site is located on a center in terms of the strength of the sources or sinks in the region. But the assumption might be reasonable enough for the overall estimate of the magnitude in consideration of other sources of uncertainty.

Finally, only one scenario under which the constraint can apply is identified in this study. Other scenarios under which either the same constraint may apply or the water vapor measurements constrain the effects of the horizontal advection of CO$_2$ in different ways are not identified.

The limitations restrict the application of the water vapor constraint in practice. For more accurate evaluation or reducing the estimated range, more constraints or direct measurements are needed.
5.6 Summary

The effects of horizontal advection on the estimates of regional surface CO$_2$ fluxes are evaluated on the spatial scale of 20km using aircraft data during IHOP 2002, indicating that ignoring such effects can lead to significant errors depending on the flow structure and surface heterogeneity. For the case of the west track with the small surface flux and heterogeneous surface conditions, the effects of the horizontal advection can be of the order of the surface flux. In contrast, the effects are smaller over the central track with the more homogeneous surface conditions. The analyses in turn indicate that the effects of the horizontal advection can be significant on the spatial and temporal scales of the order of $10^2$km and one hour, respectively, where the CBL budget method usually applies to infer regional fluxes.

As an effort to assess the effects of the horizontal advection of CO$_2$, we compared the sums of two terms in the budget equation, i.e., time rate of change term and horizontal advection term, for CO$_2$ and water vapor mixing ratios, and then hypothesized that the ratio of the sums for the two scalars is proportional to that of the corresponding surface fluxes. The analyses of the aircraft data suggest that the proportion can be confined in a range between 0 and 1 in the afternoon over highly vegetated areas. In contrast, over little vegetated areas, the proportion is much more scattered. According to those results, we made a further hypothesis that the proportion is possibly confined between 0 and 1 over vegetated areas in the afternoon under fair weather conditions. This hypothesis can be employed to constrain the horizontal gradients of CO$_2$ mixing ratio given the measurements of water vapor mixing ratio in practice.
With the conservation equations, a scenario is identified in the atmosphere to interpret the phenomenon. Under most fair weather conditions, the water vapor constraint approach provides a lower bound for the horizontal advective fluxes of CO₂ over highly vegetated areas in the afternoon during the growing season mainly due to the feature of the evolution of $WUE$ with time.

The water vapor constraint approach is employed to examine the effects of the horizontal advection on the regional estimates of the CO₂ surface fluxes over a forested area in northern Wisconsin. A lower bound for the horizontal advective fluxes suggests that ignoring the contribution of the horizontal advection could likely overestimate and underestimate the magnitudes of the surface fluxes when winds are from the NE, E, and SE directions and when wind is from the NW direction, respectively. With respect to other wind directions, the signs of the effects cannot be determined by the lower bound alone with the upper bound estimated from a scaling argument due to insufficient information. To further constrain the horizontal advective fluxes, the effects of the horizontal advection are assumed to have the same magnitudes and reversed signs in the opposite wind directions, which could be fairly reasonable based on the spatial distribution of the water vapor mixing ratio. Combination of the results estimated under this assumption and with the water vapor constraint suggests that the effects of the horizontal advection are likely negative with the winds come from W and SW directions, ranging approximately from -3 to 0 µmol m⁻²s⁻¹. When the winds blow from the N and S directions, the horizontal advective fluxes are likely between -2 and 2 µmol m⁻²s⁻¹. Overall, the magnitude of the horizontal advective flux is most likely smaller than about 15-20% of that of the surface flux in each wind direction.
The limitations and uncertainty of the application of the water vapor constraint are discussed, somewhat restricting its application in the atmosphere. For more accurate evaluation, more constraints or direct measurements are needed.
Figure 5.1: Schematic of locations of surface stations (filled circles) and UWKA flight tracks (solid lines) in the IHOP_2002 experiment. Approximate mean NDVI, types of surface vegetations, and flight dates are shown for each track.
Figure 5.2: Comparison of surface fluxes and horizontal advective fluxes for water vapor and CO₂ on a spatial scale of 20km, (a) west track with NDVI=0.1 on May 29, 2002; (b) central track with NDVI=0.45 on June 16, 2002. Water vapor flux is scaled by the right axis. The error bars are the standard deviations of the means.
Figure 5.3: Comparison of WUE and the ratio of the sums of the rate of change and horizontal advection terms for CO$_2$ and water vapor mixing ratios. Error bar is the standard deviation of the mean. The coefficient, $\alpha$, in Eq. (5.3) can be evaluated by the slope of the line linking the data point with the origin. Results from different flights and times are indicated by symbols and colors, respectively.
Figure 5.4: Evolution of the vertical profiles of (a) CO$_2$, and (b) water vapor mixing ratios with time on May 18, 2001. Data are collected near the WLEF tower area on a powered parachute platform (Schulz et al., 2004).
Figure 5.5: Evolution of the mean $WUE$, measured over a forested area in Northern Wisconsin at the WLEF tower, with time from morning to middle afternoon in the growing season, May through September of 1999.
Figure 5.6: An example of CBL height derived from a 915MHz radar near the WLEF tower, varying with time on June 17, 1999.
Figure 5.7: Lower and upper bounds for the horizontal advective flux in each wind direction sector. Error bar is the standard deviation of the mean. The lower and upper bounds estimated with the water vapor constraint are plotted by filled circles and open squares, respectively. The dot dash and dashed lines represent the lower and upper bounds estimated with the assumption that the horizontal advective fluxes have the same magnitudes and reversed signs in the opposite directions.
Chapter 6

Comparison of Regional NEE Estimates

Figure 6.1 compares the regional NEE estimated from the ABL budget method (Chapter 4) and flux aggregation (Chapter 3) in this study and from other studies (Desai et al., submitted; Helliker et al., 2004) reported in the literature. Desai et al. (submitted) conducted a similar CO$_2$ flux aggregation experiment in the same region as this study using the same vegetation map, where, different from this study, NEE for each ecosystem type is estimated from the measurements at towers located in a much larger region, the upper Midwest region. Helliker et al. (2004) applied quasi-equilibrium concepts to estimating the regional net flux of CO$_2$ in 2000 based on ABL-to-free-troposphere CO$_2$ and water vapor differences. According to their arguments, the spatial scale of their estimates is about $10^4$ to $10^5$ km$^2$, which is larger than that of this study.

During the dormant season, the regional NEE values estimated from the ABL budget method and measured at the WLEF tower are in close agreement. Particularly, they are close to the NEE value estimated by Helliker et al. (2004) on the larger scale in 2000, suggesting that the measurements at the WLEF tower can be a reasonable estimate of NEE not only in the region of this study but also in a larger region ($10^4$ to $10^5$ km$^2$). Measurements at the WC and LC sites likely overestimate and underestimate the regional NEE, respectively, indicating that the distinction between wetland and upland ecosystems might be still necessary in the season. The small difference among the three NEE values...
measured at the WLEF tower, estimated from the ABL budget calculation, and estimated by aggregating measurements at the WC and LC towers, however, suggests that the classification scheme based on watershed function might be sufficient for flux aggregation in the region during the season because CO$_2$ exchange is relatively low for all ecosystems.

During the growing season, NEE measured at the WLEF tower most likely underestimates the net uptake of CO$_2$ in the region, compared with those regional estimates. The aggregated NEE from the measurements at the LC and WC sites is more negative than other regional estimates, implying that the regional net uptake of CO$_2$ would be overestimated if the measurements at both tower sites were assumed to be representative of the respective ecosystem types in the region. The NEE value estimated from the budget calculation agrees well with that (method 4, in Figure 6.1) estimated by aggregating the fluxes both decomposed in the WLEF footprint area and measured at the WC and LC sites as well as that reported in Desai et al. (submitted). The NEE value estimated by extending the decomposed fluxes in the WLEF footprint area to the region could still underestimate regional NEE compared with that inferred from the budget calculation, suggesting that the NEE values for ecosystem types in the WLEF footprint area may not be representative of the region as analyzed in Chapter 3. In addition, the NEE value estimated on the larger scale (Helliker et al., 2004) is smaller than those estimated in the 40×40 km$^2$ region. The differences among the regional NEE estimates and tower measurements, therefore, suggest that it is inappropriate to simply extend NEE measurements or estimates on smaller scales to larger scales, e.g., from 1km$^2$ (single tower measurement) to 40×40 km$^2$ (budget calculation), and from 40×40 km$^2$ to a larger
region (Helliker et al., 2004). These comparisons also indicate that the responses of various ecosystem types to similar environmental conditions are significantly different in this heterogeneous region and that more measurements and robust modeling are needed to obtain more accurate regional estimates.

By combining the results from the top-down (method 1) and bottom-up (method 4) methods (Figure 6.1), the seasonally-averaged daily regional NEE values are estimated to be about -1.70±0.31 g C m⁻² and -1.93±0.38 g C m⁻² during the growing season in 2000 and 2003, respectively, which are significantly larger in magnitude than those measured at the WLEF site and smaller than those at the WC site. In 2003, the regional NEE is close to that measured at the LC tower (-1.53±0.16 gC m⁻²). During the dormant season, the measurement at the WLEF tower, which are about 0.53±0.03 gCm⁻² and 0.60±0.05 gCm⁻² in 2000 and 2003, respectively, can be reasonable approximations to NEE in the region, even in a larger region.

The annual cumulative regional NEE values are about -103±50 gC m⁻² and -175±60 gC m⁻² in 2000 and 2003, respectively, which are significantly different from the measurements individually at the three towers. In the calculation, NEE values are approximated as those from the above combined estimates (methods 1 and 4) and from the ABL budget method during the growing and dormant seasons, respectively. The regional estimates indicate that the ecosystem in the region is a net sink of CO₂, which is contrary to the suggestion from the data directly measured at the WLEF tower if we had assumed the measurements at the WLEF tower over the mixed ecosystem were representative of the region.
Figure 6.1: Comparison of seasonally-averaged daily-integrated regional NEE estimates from different methods. Error bar is the standard deviation of the mean. Results in Method 1 are from ABL budget calculation; Method 2: aggregated NEE from the measurements at WC and LC sites. Method 3: aggregated NEE from decomposed fluxes in the WLEF footprint area; Method 4: same as Method 2 except that WC and LC fluxes are used for aspen-excluded deciduous forests and lowland wetlands; Method 5: Multi-tower aggregation (Desai et al., submitted); Method 6: application of ABL quasi-equilibrium concept (Helliker et al., 2004).
Chapter 7

Conclusions and Future Work

7.1 Summary and conclusions

The net ecosystem-atmosphere exchange of CO₂ is estimated over a mixed forest ecosystem in the 40×40km² region centered at the WLEF tower in northern Wisconsin. Flux aggregation and the ABL budget calculation methods are employed, a bottom-up method and a top-down method, respectively.

In the bottom-up method (Chapter 3), CO₂ fluxes for six more ecosystem types are inferred in the WLEF footprint area by decomposing eddy-covariance fluxes measured at the tower using footprint models and ecosystem models. Then the fluxes, in addition to measurements at two short towers (WC and LC towers), are used for flux aggregation in the region. The flux decomposition is helpful to interpret NEE measurements at the WELF tower whose footprint areas are covered by mixed stands. To simulate the footprints of flux measurements above the surface layer, an approximate analytical model is developed by adjusting the analytical solutions for the footprint function derived in an idealized CBL to those from a stochastic model with more realistic meteorological conditions (Chapter 2). The model can be used in the lower part of the mixed layer (0.1h to 0.6h), and in the stability range of –L/h between 0.01 and 0.1, where
$L$ is Monin-Obukhov length and $h$ is the depth of the CBL, and in the range of the roughness length between $10^{-5}$ and $2 \times 10^{-3} h$. Footprints for flux measurements within the surface layer are estimated using an existing model reported in the literature. The influences of surface heterogeneity on footprint predictions are ignored, which might be a significant source of uncertainty in the analysis and should be considered in further studies. Two simple CO$_2$ flux models are used in the analysis, a light response model for daytime fluxes, and an exponential model for nighttime fluxes. The flux footprint models, a vegetation map, and eddy-covariance data at three levels of the WLEF tower are combined to estimate the optimal solutions for the functional parameters for each stand type by solving a set of nonlinear equations. The results show differences in terms of the functional parameters and fluxes among the different stand types that are consistent with general expectations for the respective stand types. Major conclusions from the analyses are as follows.

(1) From the footprint perspective, NEE measured at the WLEF tower (particularly at the lowest level) might not be representative of the region by comparing the weights of the stand types to the measured flux in the footprint area with the fractional coverage of the corresponding stand types in the region. It is hard to interpret the temporally-integrated NEE over the heterogeneous ecosystem due to time-dependent footprint functions. Preliminary analyses imply that the measured NEE integrated over a day or longer may be overestimated mathematically at the WLEF tower.

(2) A simple up-scaling hypothesis that the fluxes measured at the WLEF tower are equal to the area-weighted average of the fluxes measured at the WC and LC
towers is rejected, implying that more detailed ecosystem classification schemes are needed. The respiration rates of stand types are generally higher in the WLEF footprint area than those of the same stand types at the WC and LC sites, while GEP values for the upland forests are lower in the WLEF footprint area than at the WC site. The differences imply that the six-stand classification scheme does not capture all the variability in stand characteristics relevant to CO₂ exchange. In addition, the comparison among the decomposed fluxes implies that forested wetland and aspen ecosystems have high ER and GEP, which might contribute in part to the high ER measured at the WLEF tower. The varying fluxes of the same stand type with location challenge the widely-used land-cover based ecosystem classification scheme, where other factors such as stand age, density, and forest understory that also control the carbon exchange cannot be taken into account. More detailed schemes are needed for more accurate flux aggregation.

(3) It is critical for flux aggregation to distinguish the wetland from the upland. The difference in cumulative NEE values measured at the WC and LC site can be as large as 400 gC m⁻² season⁻¹ over the growing season. Further, fluxes among three upland forests and between forested and lowland wetlands are different in the WLEF footprint area, suggesting that a distinction among them is also important for flux aggregation. The difference in aggregated NEE values with the watershed function classification scheme and with the stand-type level classification scheme can reach about 250 gC m⁻² season⁻¹ over the entire growing season in 2003.

Regional fluxes are also estimated by the ABL budget technique, an independent approach. We take advantage of the vertical profile of CO₂ mixing ratios measured on the
tall tower and long-term study on the boundary layer in the region to infer regional NEE by calculating the budgets in the conservation equation of CO₂ under fair weather conditions (Chapter 4).

In the day, CO₂ mixing ratio is assumed to vary linearly with height above the CBL to a height of 3-4 km. With this assumption, CO₂ mixing ratios above the CBL are estimated by a combination of measurements within the CBL and in the lower troposphere, showing that CO₂ jump at the top of CBL varies seasonally and diurnally. Monthly diurnally-averaged terms in the budget equation are estimated, indicating that the magnitude of the entrainment flux can be generally more than half of that of the surface flux. The magnitude of the time rate of change term is smaller than the entrainment flux term in the developing CBL during the growing season, while both terms have similar magnitudes but opposite signs in the dormant season. The estimated NEE values in the day are compared with those measured at the three levels of the tall tower by the eddy-covariance method, suggesting that NEE measured at higher levels of the tower is closer to the regional estimates. The effects of horizontal advection on inferring the regional flux are estimated by the water vapor constraint approach (Chapter 5). With the spatial distribution of the measured water vapor mixing ratio, the magnitude of the horizontal advective flux of CO₂ is estimated to be smaller than 15-20% of the surface flux on the scale of 40km. The horizontal advective effects could be positive when the winds are from the NE, N, and NW directions but negative when the winds are from the SW, S, and SE directions. In other wind directions, the signs of the advection effects cannot be determined.
In the nocturnal boundary layer, criteria are defined to remove the cases when systematic errors or non-representative sampling occurs. The cases when the atmosphere is under very stable ($u_*<0.2\text{m/s}$) or nearly-neutral ($u_*>0.6\text{m/s}$) conditions are screened out. About 40% of data are discarded. Uncertainty due to the use of the vertical profile of CO$_2$ mixing ratios measured at one location is discussed. The relative error can be as large as 35% under very stable conditions but only 15% under well-mixed conditions ($u_*>0.2\text{m/s}$). The advection terms are ignored in the evaluation, which can lead to significant errors on short time scales but can possibly be reduced on long time scales. Uncertainties due to such ignorance cannot be determined without measurements.

All estimated NEE values are fitted to the light response model or the exponential function of air temperature. The regional daytime NEE is generally bounded by current tower measurements. During the growing season, the daytime NEE at the WC site is the most negative, with the least being observed at 30m of the WLEF tower. The largest nighttime NEE is observed at the WLEF tower, followed by the regional NEE estimate. During the dormant season, both nighttime and daytime NEE are the largest at the WC site. In the day, NEE measured at 30m of the WLEF tower is the smallest. The same is true of the NEE measured at the LC site at night.

The inferred regional NEE values from aggregation and ABL budget methods are compared with each other and with the results reported in the literature. The comparison suggests that the measurements at the WLEF tower can be a reasonable approximation of NEE in the region, or even a larger region ($10^4$ to $10^5 \text{km}^2$) during the dormant season because CO$_2$ exchange rates and their differences among ecosystems are relatively low. This, however, is not the case during the growing season. The regional NEE estimates
differ significantly from those measured at the towers. The averaged daily regional NEE values are estimated to be about $-1.70 \pm 0.31$ g C m$^{-2}$ and $-1.93 \pm 0.38$ g C m$^{-2}$ in 2000 and 2003, respectively, whose magnitudes are significantly larger than those measured at the WLEF tower and smaller than those at the Willow Creek tower. Annually, measurements indicate that the ecosystem in the WLEF footprint area is a net source of CO$_2$ (e.g., annual cumulative NEE is about $111 \pm 16$ g C m$^{-2}$ in 2000 (Ricciuto et al., submitted)). In contrast, the ecosystem in the WC footprint area is a large net sink of CO$_2$ (e.g., annual cumulative NEE is about $-334$ g C m$^{-2}$ in 2003 (Cook et al., 2004)). In this study, the annual cumulative regional NEE values are estimated to be about $-103 \pm 50$ gC m$^{-2}$ and $-175 \pm 60$ gC m$^{-2}$ in 2000 and 2003, suggesting that the region is a moderate net sink of CO$_2$. It is inappropriate simply to extend fluxes measured at individual towers to a larger region.

7.2 Future work

Limitations are also discussed for each method in addition to the developed footprint model and the water vapor constraint approach in the dissertation. Future efforts that can be made to improve this study include the following:

1. Although an approximate analytical footprint is developed for the measurements above the surface layer, the assumption that the atmospheric conditions are homogeneous horizontally still restricts the application of the model as well as other existing analytical models to accurately predicting footprints in practice
over inhomogeneous surfaces and under complex atmospheric conditions. The performance of the flux decomposition method described in this study relies largely on the footprint predictions. As a result, developing advanced footprint models that both can handle the complex flow and terrain and can be run efficiently is a priority to apply the flux decomposition method.

(2) More accurate classification data are needed to conduct the flux decomposition over heterogeneous ecosystems based on the interpretation of footprints. Besides, more detailed classification schemes that can take into account other controlling factors such as stand age and forest floor features are needed for flux aggregation in the region with heterogeneous ecosystems like that in this study.

(3) To constrain the inferred functional parameters for the ecosystem types, measurements for at least one stand type or knowledge of the relationship of the parameters or fluxes among stand types is needed. In doing so, the optimal solutions for the functional parameters can be determined by conducting numerical experiments and judging the results when other uncertainties exist and cannot be quantified with sufficient confidence, e.g., in footprint predictions and the ecosystem map. In addition, quantifying uncertainties in the footprint predictions and flux measurements is also critical to obtaining more precise and accurate solutions.

(4) As for the application of the ABL budget method, developing models that can simulate the CO$_2$ jump as a function of variables routinely measured can improve budget calculations. In addition, the estimate of the vertical velocity at the CBL top needs to be improved either by models or by measurements.
(5) It is still difficult to evaluate the effects of the horizontal advection on the surface flux estimates. More constraints from the measurements of other tracers or developing other approaches is needed to further narrow the range of the estimates of the effects of the horizontal advection when CO₂ mixing ratios are measured only at one location. At night, measurements at more than one location are needed because of the localized distribution of CO₂ in the stable atmosphere. Eventually, strategies of measuring horizontal distribution of CO₂ mixing ratio should be developed.

(6) Due to the assumptions used and ignorance of some terms in the budget equation, regional NEE can be estimated with more confidence only on longer time scales such as months and seasons. To study the mechanisms of the response of ecosystem to environmental conditions on regional scales and develop process-based models, more experiments need to be designed to measure NEE on various temporal scales from hours to years. In addition, regional estimates can be possibly biased as a result of the assumptions used. It would be better to compare the results independently from more approaches.


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Appendix A

Derivations and Coefficients in the Fitted Curves in Chapter 2

A.1 Expression of the vertical turbulent flux

Following the derivation by van Dop et al. (1985), we can write the mean vertical flux of the mixing ratio of a scalar in a fluid with zero mean vertical flow in the Eulerian field as, by definition,

$$
\overline{w c}(x, y, z) = \frac{1}{N} \sum_{i=1}^{N} \tilde{w}_i(x, y, z) \tilde{c}_i(x, y, z),
$$

(A.1)

where $N$ is the total number of the fluid particles passing though an infinitesimal volume centered at point $(x, y, z)$; $\tilde{w}_i$ and $\tilde{c}_i$ are instantaneous values of vertical velocity and mixing ratio, respectively. Of the particles, only a subensemble of $M$ ($M \leq N$) particles has passed through the source $z_s$ with a constant mixing ratio $q$, and hence the mixing ratio at the point is that $C(x, y, z) = qM/N$. Since $\tilde{c}_i$ is only nonzero (equal to $q$) for the particles having passed the source, Eq. (A.1) reduces to, Eq. (A.2)

$$
\overline{w c}(x, y, z) = \frac{q}{N} \sum_{i=1}^{M} \tilde{w}_i(x, y, z) = q \frac{M}{N} \frac{1}{M} \sum_{i=1}^{M} \tilde{w}_i(x, y, z),
$$

(A.2)

$$
= \langle w(x, y, z) \rangle C(x, y, z)
$$
where \( <w(x, y, z)> \) is the mean vertical velocity of the \( M \) particles. Eq. (A.2) was shown by van Dop et al. (1985).

In the two-dimensional case, Eq. (A.2) can be reduced to,

\[
\overline{wc}(x,z) = \frac{q}{N} \sum_{i=1}^{M'} \tilde{w}_i(x,z) = \frac{q}{M'} \frac{1}{M'} \sum_{i=1}^{M'} \tilde{w}_i(x,z) = <w(x,z)> C(x,z), \quad (A.3)
\]

where \( <w(x,z)> \) is the mean vertical velocity of the \( M' \) (obviously, \( M' \geq M \)) particles passing through the source and point \((x, z)\); \( C(x, z) \) is the mixing ratio at point \((x, z)\), which is equivalent to the y-direction integrated mixing ratio and usually written as \( C^y(x,z) \).

For the special case in section 2.3 (chapter 2) where the vertical velocities of the particles passing the source can be explicitly expressed as \( n \) discrete values such as \( \tilde{w}_1, \tilde{w}_2, \ldots, \tilde{w}_n \), Eq. (A.2) or (A.3) can be rewritten as,

\[
\overline{wc} = \frac{q}{N} \sum_{i=1}^{M} \tilde{w}_i = \frac{q}{N} \sum_{i=1}^{n} n_i \tilde{w}_i = \sum_{i=1}^{n} \frac{q}{N} n_i \tilde{w}_i = \sum_{i=1}^{n} C_i \tilde{w}_i, \quad (A.4)
\]

where \( C_i \) and \( C_i \tilde{w}_i \) are the mixing ratio and the vertical flux, respectively, contributed by the \( n_i \) \( (\sum_{i=1}^{n} n_i = M) \) particles with the vertical velocity of \( \tilde{w}_i \). Eq. (A.4) yields the result, i.e., Eq. (2.21).

### A.2 Coefficients in the fitted curves

The Dimensionless \( AX_{\text{stochastic}}, F_{\text{max, stochastic}}, \) and \( X_{\text{max, stochastic}} \) are fitted as Eqs. (2.14), (2.15), and (2.16), where the coefficients are dependent on the stability \((L/h)\).
and roughness length \( (z_0/h) \). The coefficients are fitted using the following polynomial function,

\[
g_k \left( \frac{L}{h}, \frac{z_0}{h} \right) = a_k(0) + a_k(1) \frac{L}{h} + a_k(2) \frac{z_0}{h} + a_k(3) \frac{L z_0}{h^2} + a_k(4) \left( \frac{L}{h} \right)^2 + a_k(5) \left( \frac{z_0}{h} \right)^2,
\]

(A.5)

where \( g_k \) represents the \( k \)th coefficient in the equations, \( k \) is from 1 to 6; the values of \( a_k \) (0:5) are the fitted coefficients for \( g_k \). The coefficients of determination \( (r^2) \) for the fitted curves are greater than 0.95.

For Eq. (2.14), \( \Delta X_{\text{stochastic}} = g_1 \left( \frac{z}{h} \right)^{z_2} \),

\( a_1=\left[ 0.41, -7.07, 12.53, 959.65, -30.45, 0.00 \right]; \)

\( a_2=\left[ 0.67, 5.08, 97.26, -62.77, 0.00, 0.00 \right]. \)

For Eq. (2.15), \( F_{\text{max, stochastic}} = g_3 \left( \frac{z}{h} \right)^{z_4} + g' \left( \frac{z}{h} \right)^{z_5} \),

\( a_3=\left[ 1.33, 11.24, 11.53, -500.30, 7.89, 0.00 \right]; \)

\( a_4=\left[ -0.69, -4.78, -45.59, 679.85, 0.00, 0.00 \right], \) and \( g'=-0.95. \)

For Eq. (2.16), \( X_{\text{max, stochastic}} = g_5 \left( \frac{z}{h} \right)^{z_6} \),

\( a_5=\left[ 0.61, -1.97, -9.80, 494.18, 7.89, 0.00 \right]; \)

\( a_6=\left[ 0.78, 2.45, 13.44, -915.30, 0.00, 0.00 \right]. \)
With the parameters used ($R=1$, $S=0.5$) in this study, $\Delta X_{\text{ideal}}, X_{\text{max, ideal}},$ and $F_{\text{max, ideal}}$ for the idealized model, i.e., Eq. (2.22) (Figure 2.5) can be fitted by the following curves,

$$\Delta X_{\text{ideal}} = \sum_{i=1}^{5} d_i \left(\frac{z}{h}\right)^i$$  \hspace{1cm} (A.6)

where $d_i$ ($i=1$ to 5) is the coefficient of the fitted polynomial function; $d=[1.131, 5.7192, -17.822, 17.058, -5.4377]$;

$$X_{\text{max, ideal}} = 1.2305 \left(\frac{z}{h}\right)^{0.993}, \text{ for } z < 0.6h,$$ \hspace{1cm} (A.7)

and,

$$f_{\text{max, ideal}} = 0.3766 \left(\frac{z}{h}\right)^{-1.0028}, \text{ for } z < 0.6h,$$ \hspace{1cm} (A.8)

The coefficients of determination ($r^2$) for the fitted curves are greater than 0.99.

### A.3 Examining Eq. (2.23)

The condition Eq. (2.23) is examined as follows:

$$\int_{0}^{+\infty} f_{a}^{\gamma}(x, z_{m}) \, dx = \int_{0}^{+\infty} \beta f^{\gamma}(\beta x + \gamma, z_{m}) \, dx,$$ \hspace{1cm} (A.9)

Let

$$x' = \beta x + \gamma,$$ \hspace{1cm} (A.10)

Eq. (A.9) becomes,
The second integral on the far right-hand side of Eq. (A.11) is equal to zero if \( \gamma \leq 0 \) according to Eq. (2.22). In this case, the constraint Eq. (2.23) is satisfied. If \( \gamma > 0 \), the second integral is nonzero. For a small \( \gamma \), the integral can be negligible because the footprint near the source is close to zero (e.g., Figure 2.6). Calculation indicates that the integral is significant only when \( z_m \) is small (e.g., less than 0.02\( h \)). In this case, the position of the footprint peak is close to the receptor. For the cases above the surface layer, the integral is negligible. Therefore, the constraint Eq. (2.23) is approximately satisfied for the range of height where the model is applicable.

The case of \( \gamma > 0 \) indicates that surface sources downwind of the receptor (\( x < 0 \), note that the wind comes from right to left) could contribute to the measured flux. In particular for the measurements close to the surface under extremely unstable conditions, such contribution could be significant due to the effect of the along-wind turbulence, which is also shown by numerical studies (Kljun et al., 2002). In this case, with the inclusion of the footprint at negative \( x \), the integral in the constraint Eq. (2.23) is rewritten as,

\[
\int_{-\infty}^{+\infty} f_a^\gamma (x, z_m)dx = \int_{-\infty}^{+\infty} f^\gamma (x, z_m)dx = 1 - \frac{z_m}{h}, \tag{A.12}
\]

and the adjusted model Eq. (2.24) can be rewritten accordingly as,
where the contribution of the surface sources downwind of the receptor is considered. In this case, Eq. (A.13) exactly meets Eq. (A.12).
Appendix B

Estimating Effects of the Interfacial Layer on Estimating Surface Fluxes

Two terms in the budget equation, i.e., the time rate of change term and the entrainment flux term (see Eq. 5.6), can be affected by considering the interfacial layer at the top of CBL in the calculation due to changes in the mean CO₂ mixing ratio. The depth of the interfacial layer is estimated in two ways. One is to assume that the depth is simply equal to a fraction of the CBL depth. The other is to compute the depth as a function of environmental variables according to a model fitted to observational data.

Field observations (e.g., Beyrich and Gryning, 1998; Boers, 1989; Davis et al., 1997) suggest that the depth of the interfacial layer is typically 20% of the CBL depth, which is used to examine its impacts on the surface flux estimates. Two additional assumptions are made in the evaluation. First, the mean height of the CBL top is in the middle of the interfacial layer, i.e., the bottom and top of the interfacial layer (Figure 5.1) are located at \( h_1 = h - 0.5 \Delta h \) and \( h_2 = h + 0.5 \Delta h \), respectively, where \( \Delta h = h_2 - h_1 \) is the depth of the interfacial layer. Second, CO₂ mixing ratio is distributed linearly with height to facilitate calculating the mean mixing ratio in the CBL. With the interfacial layer being considered, the monthly average of the time rate of change term increases and decreases by approximately 10% during the growing and dormant seasons (Figure B.1a), respectively, compared to those without the interfacial layer being considered as shown in section 5.4. In contrast, the monthly average of the entrainment flux term decreases
and increases by roughly 5% and 10% during the growing and dormant seasons, respectively. Therefore, the changes in both terms are in part compensated, resulting in a small increase in the surface flux by 1% and 5% during the growing and dormant seasons, respectively. Even with the interfacial layer depth being doubled, the relative change in the surface flux is still smaller than 5%, though the changes of the two terms increase individually by a factor of about 1.5 to 2.

More realistically, the interfacial depth can be described as a function of the convective Richardson number based on lidar observations and lab experiments. For example, the model of the normalized entrainment depth fitted to lidar observation by Boers (1989) reads,

\[
\frac{\Delta h}{h} = 1.38Ri^{-0.52} \tag{B.1}
\]

where \( Ri \) is the convective Richardson number that is defined as,

\[
Ri = \frac{gh\Delta \theta}{\theta_0 w^*} \tag{B.2}
\]

where \( g \) is the gravity; \( w^* \) is the convective velocity scale; \( \theta_0 \) is the mean potential temperature in the boundary layer; \( \Delta \theta \) is the virtual potential temperature jump at the inversion interface. By assuming that the potential temperature varies linearly with height within the interfacial layer, i.e., between \( h_1 \) and \( h_2 \), and above the top of the interfacial layer, i.e., \( h_2 \), respectively, the jump \( \Delta \theta \) can be expressed as,

\[
\Delta \theta = \theta(h_2) - \theta(h_1) = \Gamma \times (h + \lambda \Delta h) + \theta_s - \theta_0 \tag{B.3}
\]
where $\Gamma$ is the slope of the vertical gradient of the potential temperature with height above $h_2$; $\lambda$ is a fraction determining the position of the mean CBL top between $h_1$ and $h_2$, i.e., $h = \lambda h_1 + (1- \lambda) h_2$, which is 0.5 on the assumption that the mean CBL top is at the middle of the interfacial layer; $\theta_s$ is the potential temperature when the potential temperature profile is extended to the surface. The potential temperature profile above the CBL (far from $h_2$) can be obtained from the reanalysis data or radiosonde sounding data near the site. The values of $w^*$ and $\theta_0$ are estimated from measurements at the tall tower. Therefore, with Eq. (B.3) being substituted into Eq. (B.2), the interfacial layer depth can be evaluated by solving a nonlinear equation, which is about 15-20% of the mean depth of the CBL.

Similarly, the calculation shows that the effects of the interfacial layer on the time rate of change term and entrainment flux term can be significant respectively, but they are inversely correlated, leading to a small impact on their sum, i.e., the estimated surface flux (Figure B.1b).
Figure B.1: The monthly average differences in time change rate, entrainment flux, and surface flux terms in the budget equation with and without the interfacial being considered. In (a), the interfacial layer depth is assumed to be 20% of the mean height of the CBL top according to recent field observations. In (b), the interfacial layer depth is predicted using an analytical model derived from lidar data, in which the depth changes in response to environmental variables such as Richardson number near the top of CBL. For comparison, typical surface fluxes are about -10 and 1 \( \mu \text{mol m}^{-2}\text{s}^{-1} \) in the growing season and dormant season, respectively. The error bars are the standard deviations of the means.
Appendix C

Evaluating the Ratio of $k_c$ and $k_q$

C.1 Evaluation of Eq. (5.17)

The budget equation for CO$_2$ without the advection terms being considered, Eq. (5.11), can be rewritten as,

$$ F_c = h \frac{\partial [c]_m}{\partial t} - \left( \frac{\partial h}{\partial t} \right) ([c(h)]_+ - [c]_m) $$

$$ = h \frac{\partial [c(h)]_+}{\partial t} - h \frac{\partial ([c(h)]_+ - [c]_m)}{\partial t} - \left( \frac{\partial h}{\partial t} \right) ([c(h)]_+ - [c]_m) \cdot \quad \text{(C.1)} $$

$$ = h \frac{\partial [c(h)]_+}{\partial t} - \frac{d}{dt} \{ h(t)([c(h)]_+ - [c]_m) \} $$

Provided that the effects of the horizontal advection is negligible above the CBL compared to those of the vertical advection, the change in $[c(h)]_+$ with time can be formulated as (Tennekes, 1973),

$$ \frac{\partial [c(h)]_+}{\partial t} = \frac{\partial [c(z)]}{\partial z} \bigg|_{z=h} \times \left( \frac{\partial h}{\partial t} - w_z \right) = \Gamma_c(h) \left( \frac{\partial h}{\partial t} - w_z \right), \quad \text{(C.2)} $$

where $[c(z)]$ is the CO$_2$ mixing ratio above the CBL; $\Gamma_c(h)$ is its vertical gradient at height $h$. If we further assume that $\Gamma_c$ is constant with height and time, i.e., the CO$_2$ mixing ratio is linearly distributed with height above the CBL, substituting Eq. (C.2) into Eq. (C.1) yields,
Integrating Eq. (C.3) from 
\[ t_{c0} \] to 
\[ t \], we have,

\[
F_c = \frac{d}{dt} \left\{ -h([c(h)]_n - [c]_m) + \frac{1}{2} h^2 \Gamma_c \right\} - \Gamma_c w_c h .
\]  \hspace{1cm} (C.3)

If \( t_{c0} \) is chosen as the time when \([c]_m = [c(h)]_n\), substituting \( I_c \) into Eq. (C.4) yields Eq. (5.12). Eq. (C.4) is reduced to the expression shown by Raupach et al. (1992) under the assumption that either the CO\(_2\) mixing ratio above the CBL is constant or the magnitude of the change in the \([c(h)]_n\), with height is small compared to that of \(( [c(h)]_n - [c]_m)/h )/ h.

Similarly, with respect to water vapor, we can have,

\[
\int_{t_{c0}}^{t} F_q dt = \left\{ -h(t)([q]_n - [q]_m) + \frac{1}{2} \Gamma_q h(t)^2 \right\} - \Gamma_q \int_{t_{c0}}^{t} w_q h(t) dt ,
\]  \hspace{1cm} (C.4)

where the water vapor mixing ratio above the CBL is assumed to invariant with time and linearly distributed with height; If \( t_{q0} \) is chosen as the time when \([q]_m = [q(h)]_n\), substituting \( I_q \) into Eq. (C.5) yields Eq. (5.13). The selection of the initial times, rather than the time when \( h=0\), is because it is more reasonable to assume that the CO\(_2\) and water vapor mixing ratios vary linearly with height above the CBL after \( t_{c0} \) and \( t_{q0} \) in the middle day, respectively, when the CBL top is not too low. Moreover, the magnitude of the WUE is observed typically decreasing with time from the time when \([c(h)]_n = [c]_m\) to late afternoon, convenient to interpret the ratio of \( k_c \) and \( k_q \). The initial time \( t_{c0} \) can be
usually found between early and middle morning when plants assimilate the respired CO₂ accumulated in the nocturnal boundary layer. In contrast, the time of \([q]_r=\[q\]_m\) can not be always found, e.g. when air is very dry aloft. In this case, the initial time when \(h=0\) can be used as \(t_{q0}\) instead for interpretation.

The two assumptions in deriving Eq. (C.4) and Eq. (C.5) can be justified in part by observations. Firstly, the CO₂ and water vapor mixing ratios above the CBL are decoupled with those within the CBL, which are heterogeneous mostly on the scales larger than those within the CBL. As a result, it is reasonable to assume that they are homogeneous on the horizontal scale we are considering. Secondly, observations indicate that the CO₂ mixing ratio above the CBL (in the middle day, at least after \(t_{c0}\)) usually increases nearly linearly with height, i.e., \(\Gamma_c>0\), over vegetated areas under fair weather conditions during the growing season (Figure 4.2) due to net uptake of CO₂ by ecosystems and daytime vigorous vertical mixing in previous days. In contrast, the CO₂ mixing ratio usually decreases nearly linearly with height during the dormant season (Figure 4.2) due to net release of CO₂ by ecosystems. With respect to water vapor, no such seasonal patterns are found. \(\Gamma_q\) is usually close to zero above the CBL in subsidence weather patterns. In addition, \(\Gamma_q\) can be greater than zero due to the impacts of clouds near the top of the CBL. Sometimes the CO₂ mixing ratio above the CBL is also observed to be nearly invariant with height after large scale disturbances in the atmosphere such as fronts and deep convections (Hurwitz et al., 2004). Note that the linear assumption also indicate the vertical gradients are independent of time under horizontally homogeneous conditions according to their conservation equations. Thus, based on those observations, typically we have,
and then,

\[
\int_{t_0}^{t} F_q(t')dt' - \Gamma_q I_q \approx \int_{t_0}^{t} F_c(t')dt',
\]

\(\text{(C.6)}\)

\[
\tau \equiv \frac{\int_{t_0}^{t} F_c(t')dt' - \Gamma_c I_c}{\int_{t_0}^{t} F_q(t')dt' - \Gamma_q I_q} \approx \frac{\int_{t_0}^{t} F_c(t')dt'}{\int_{t_0}^{t} F_c(t')dt'} \geq 1.
\]

\(\text{(C.7)}\)

Eq. (C.7) is obtained due to the opposite signs of \(F_c\) and \(\Gamma_c\) that are typically observed in the middle of the growing and dormant seasons (note that \(I_c > 0\)) when the effects of \(\Gamma_q\) is ignored. With typical values over the WLEF area in a sunny day in summer, e.g., \(F_c = 0.5\) ppm m/s, \(t_{c0} = 08:00\), \(t = 12:00\), \(h(t_{c0}) = 500\) m, and \(h(t) = 1500\) m, \(\Gamma_c = 5\) ppm/km, and \(w_+ = -0.01\) m/s, \(\tau\) is about 1.4. With zero \(\Gamma_c\) just after large scale disturbances in the atmosphere, \(\tau\) is nearly equal to 1. Therefore, Eq. (C.6) and Eq. (C.7) result in Eq. (4.17), i.e.,

\[
r(t) = \frac{k_c}{k_q} = \frac{F_c}{F_q} \tau r_w = \tau r_w \geq r_w,
\]

\(\text{(C.8)}\)

where \(r_w\) is the ratio of \(k_c\) and \(k_q\) with zero \(\Gamma_c\) and \(\Gamma_q\), i.e.,

\[
r_w(t) = \frac{F_c}{F_q} \tau r_w(t).
\]

\(\text{(C.9)}\)

With the definition of \(WUE\), Eq. (C.8) can be rewritten as,
If $t_0$ is earlier than or the same as $t_0$, i.e., $t_0 \leq t_0$, it can be readily shown that $r(t) > 1$ in the morning or middle afternoon because $WUE$ usually does not change sign and decreases with time over vegetated areas from $t_0$ to middle afternoon (Figure 4.5).

In the case of $t_0 > t_0$, the first integral in the denominator of the far right hand side of Eq. (C.10) may affect $r_w(t)$. Because the water vapor flux in the early morning is small, the impact is expected small. According to Eq. (4.13), the term at the denominator can be rewritten as,

$$WUE(t)\left\{\int_{t_0}^{t} F_q(t')dt' + \int_{t_0}^{t} F_q(t')dt'\right\} = \frac{WUE(t)\int_{t_0}^{t} F_q(t')dt'}{1 - \frac{h(t_0)\Delta[q(t_0)] - \Gamma_q I_q(t_0)}{h(t)\Delta[q(t)] - \Gamma_q I_c(t)}}. \quad (C.11)$$

Substituting Eq. (C.11) into Eq. (C.8) yields,

$$r(t) \approx \frac{\int_{t_0}^{t} WUE(t')F_q(t')dt'}{WUE(t)\int_{t_0}^{t} F_q(t')dt'}, \quad (C.12)$$

where $\bar{r}_1 = 1 - \frac{h(t_0)\Delta[q(t_0)] - \Gamma_q I_q(t_0)}{h(t)\Delta[q(t)] - \Gamma_q I_c(t)} \approx 1 - \frac{h(t_0)\Delta[q(t_0)]}{h(t)\Delta[q(t)]}$. Typically, $t_0$ is in the early morning, e.g., 07:00 or 08:00 local time. If $t$ is far later than $t_0$, e.g., at noon or early afternoon, the impact of $\bar{r}_1$ is small. Under fair weather conditions, the CBL height at
noon or early afternoon (e.g., 1500m or more) can be three times or more as large as that in the early morning (e.g., 300-500m at 07:00 or 08:00 local time) (Figure C.1b and Figure 4.6); the jump of water vapor mixing ratio in the early afternoon can be twice that in the early morning. Therefore, a typical value of \( \tau_1 \) is approximately 0.9 between noon and early afternoon with negligible \( \Gamma_q \). Therefore, with the product of \( \tau \) and \( \tau_1 \) being typically greater than unity, \( r(t) \) can be larger than unity in the late morning and afternoon due to the rapidly decreased WUE with time. As an example, Figure C.1(a and b) presents the ratios of \( k_c \) and \( k_q \) with and without \( \Gamma_c \) and \( \Gamma_q \) being considered as a function of time over the WLEF area, respectively, estimated using the monthly diurnal averages of the CBL depth (Figure C.1d), water vapor flux, and WUE in the days with fair weather, for each month during the growing season. Typically, the values of \( r(t) \) are between 1.5 and 2 in the early afternoon (Figure C.1 a); In the late morning or afternoon, \( r_w \) is greater than 1 (Figure C.1b); The factor \( \tau \) is between 1.2 and 1.6 (Figure C.1c), indicating the effects of the vertical gradients of water vapor and CO₂ mixing ratios above the CBL. In the calculation, \( \Gamma_c \) values are taken approximately as 1, 3, 4, 3, 1 ppm km⁻¹ from May to September, respectively, inferred from observations; \( \Gamma_q \) values are assumed as -0.1, -0.3, -0.4, -0.3, -0.1 g kg⁻¹ km⁻¹ from May to September, respectively. \( t_{q0} \) and \( t_{c0} \) are taken as 06:00 and 08:00 in the morning, respectively. The CBL depth is inferred from the measurements of a 915MHz boundary layer profiling radar deployed near the WLEF tower in 1998 and 1999.
C.2 Evaluation of the effects of the advection terms on Eq. (5.23)

Let $\beta_c$ and $\beta_q$ be the ratios of the advective fluxes to surface fluxes of CO$_2$ and water vapor, respectively, i.e.,

$$\beta_c = \frac{h \left[ u \frac{\partial [c]}{\partial x} \right]_m + w_c \Delta [c]}{F_c}, \quad (\text{C.13})$$

and,

$$\beta_q = \frac{h \left[ u \frac{\partial [q]}{\partial x} \right]_m + w_q \Delta [q]}{F_q}. \quad (\text{C.14})$$

Therefore, Eq. (5.23) can be rewritten as,

$$r(t) = \frac{F_q}{F_c} \left( 1 - \beta_c \right) \int_{t_0}^{t} F_c dt' - \Gamma_c I_c (t), \quad (\text{C.15})$$

where we have assumed that both ratios are invariant with time. Alternatively, both $\beta_c$ and $\beta_q$ can be interpreted roughly as the mean or median ratios during the time period of integration if the surface fluxes do not change too rapidly during the time period of integration. With Eq. (C.7) and Eq. (C.9), the above equation can be rewritten as,

$$r(t) = \frac{\tau - \beta_c}{1 - \beta_q} r_w. \quad (\text{C.16})$$
It has been shown that, without advection terms being considered, \( r(t) \) is greater than unity over the vegetated areas in the late morning and early afternoon under fair weather conditions during the growing season. Eq. (C.15) retains this feature if one of the following two conditions obtained by solving the inequality equation \( r(t) \geq 1 \) is met. They are,

1. \( \beta_q > 1 \) and \( \beta_c \geq \frac{1}{r_w} (1 - \beta_q) \);

2. \( \beta_q < 1 \) and \( \beta_c \leq \frac{1}{r_w} (1 - \beta_q) \).

It is, however, still difficult to apply the above conditions. For the sake of convenience, we derive an explicit condition that is a subset of the above conditions.

According to the scaling argument, we have,

\[
O\left(h \left[ u \frac{\partial [q]}{\partial x} \right] \right) \leq O\left(w, \Delta [q] \right) \tag{C.17}
\]

where \( O \) represents the order of the variables in the parentheses. With typical values, \( O(u)=10\text{m/s}, \ O(h)=1\text{km}, \ O(w)=0.01\text{m/s}, \) and \( O(\Delta_h[q]/\Delta[q])=0.1 \), where \( \Delta_h[q] \) is the difference of water vapor mixing ratio in the horizontal direction, the magnitude of the horizontal advection term is smaller than that of the vertical advection term if the spatial scale is of the order of \( 10^2 \text{km} \). Therefore, it is most likely that \( \beta_q \) is greater than or equal to zero under the conditions we are considering on the scale of \( 40\text{km} \) (note that \( \frac{w \Delta [q]}{F_q} \) is usually positive). In the case of the second condition, we have,

\[
\beta_c \leq \frac{\tau r_w - 1}{r_w}. \tag{C.18}
\]
With typical values of \( \tau \), e.g., \( \tau = 1.4 \), and \( \mathbf{r}_w = 1.0 \) to 1.5 (Figure C.1 b\&c) in the forested area in northern Wisconsin, the term on the right-hand side of Eq. (C.18) is between 40% and 80%, suggesting that \( r \geq 1 \) holds if the (overall) total advective flux of CO2 is smaller than 40% to 80% of the surface flux and the advective flux of water vapor is between 0 and 100% of the surface flux. The first condition is unlikely over vegetated areas in our case under the assumption of fair weather conditions and the horizontally homogeneous boundary layer.

It should be noted that Eq. (C.18) is only a sufficient condition for \( r \geq 1 \). Under other conditions, the inequality \( r \geq 1 \) may also hold.

### C.3 Derivation of Eq. (5.24)

According to Eq. (5.6) and Eq. (5.21), \( k_c \) can be expressed as,

\[
k_c(t) = \frac{-\left( \frac{\partial h}{\partial t} - w_{+} \right) \Delta[c]}{F_c(t)} = \frac{-\left( \frac{\partial h}{\partial t} - w_{+} \right) \int_{t_0}^{t} F_c' dt - \Gamma_c I_c(t)}{F_c(t)}.
\]

Then, with Eq. (C.7) and Eq. (C.13), the above equation can be rewritten as,

\[
k_c(t) = \frac{\left( \frac{\partial h}{\partial t} - w_{+} \right) (\tau - \beta_c) \int_{t_0}^{t} F_c(t) dt}{F_c(t)} - \frac{\left( \frac{\partial h}{\partial t} - w_{+} \right) (\tau - \beta_c) \Delta t}{h},
\]

where \( \tau - \beta_c \) is approximately equal to 1. For an imposed upper bound, e.g., \( k \), Eq. (C.20) yields Eq. (5.24).
Figure C.1: (a) The ratio of $k_c$ and $k_q$ as a function of time over the WLEF area in each month during the growing season when $\Gamma_q \neq 0$ and $\Gamma_c \neq 0$, calculated using monthly diurnal averages of the CBL depth, water vapor flux, and WUE in 1998 and 1999; (b) The ratio of $k_c$ and $k_q$ without the vertical gradients of water vapor and CO$_2$ mixing ratio being considered above the CBL, i.e. Eq. (C.9); (c) the factor, $\tau$, due to the effects of the vertical gradients of water vapor and CO$_2$ mixing ratio above the CBL, i.e., Eq. (C.7) with $\Gamma_q \neq 0$ (see the text) (d) the monthly diurnal average of the CBL depth that is inferred from the measurements of a 915Mhz boundary layer profiling radar deployed in 1998 and 1999.
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