Validating remotely sensed land surface fluxes in heterogeneous terrain with large aperture scintillometry
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The Large Aperture Scintillometer (LAS) has emerged as one of the best tools for quantifying areal averaged fluxes over heterogeneous land surfaces. This is particularly useful as a validation of surface energy fluxes derived from satellite sources. We examine how changes in surface source area contributing to the scintillometer and eddy covariance measurements relate to satellite derived estimates of sensible heat flux. Field data were collected on the Konza Prairie in Northeastern Kansas, included data from two eddy covariance towers: one located on an upland, relatively flat homogeneous area, and the second located in a lowland area with generally higher biomass and moisture conditions. The large aperture scintillometer spanned both the upland and lowland areas and operated with a path length of approximately 1 km specifically to compare to Moderate Resolution Imaging Spectroradiometer (MODIS) derived estimates of surface fluxes. The upland station compares well with the LAS (correlation of 0.96), with the lowland station being slightly worse (correlation of 0.84). Data from the MODIS sensor was used to compute surface fluxes using the ‘triangle’ method which combines the remotely sensed data with a soil-vegetation-atmosphere-transfer scheme and a fully developed atmospheric boundary layer model. The relative contribution to the surface observations is estimated using a simple footprint model. As wind direction varies, the relative contribution of upland and lowland sources contributing to the LAS measurements varies while the MODIS pixel contribution remains relatively constant. With the footprint model, we were able to evaluate the relationship between the LAS observations and the remotely sensed estimates of the surface energy balance. The MODIS derived sensible heat flux values correspond better to the LAS measurements (percentage error: 0.04) when there was a larger footprint compared to a time with a smaller footprint (percentage error: −0.13). Results indicate that the larger the footprint, the better the agreement between satellite and surface observations.

1. Introduction

Quantifying the spatial and temporal variability of mass and energy exchanges between the surface and the atmosphere is of vital importance for numerous reasons, including water management in the case of evapotranspiration, crop yield monitoring and assessing the impacts of regional climate change. The use of remotely sensed
data provides the best possibility for determining the spatial variability of surface energy fluxes over heterogeneous terrain. Recently, more advanced multi-sensor techniques have become available for estimating the spatial and temporal variability of surface fluxes (McCabe et al. 2008, Jimenez et al. 2009). However, these require some validation technique. While networks of eddy covariance towers are available (e.g. FLUXNET, Baldocchi et al. 2001), understanding the scaling of the flux towers to compare with a satellite measurement or compute a truly regional value is of utmost importance (Kim et al. 2006). However, the direct comparison between eddy covariance and a flux derived from a satellite pixel using the radiometric temperature, vegetation and soil moisture may not be a valid comparison (Brunsell and Gillies 2003b, Li et al. 2008). Therefore, there is a strong desire for a measurement that has a spatial resolution on the same order as those of satellite platforms in order to validate the remotely sensed estimates of fluxes.

One complicating issue when comparing the surface observations to a pixel derived estimate of the flux, is that the upwind area contributing the flux (i.e. the flux footprint) must be estimated for the observations. As wind direction and atmospheric stability varies, so will the flux footprint. Assessing the variability is not a trivial task (Schmid 2002). Over homogeneous terrain, a direct comparison between the surface and a satellite is possible. However, over heterogeneous terrain, the variation in contributing areas between the observations must be addressed (Schmid 2002).

Given the number of eddy covariance stations located around the globe with continuous monitoring of surface-atmosphere exchange, it is only reasonable that the focus on verification of satellite fluxes be on the comparison with these tower measurements. In cases where high spatial resolution imagery is available, a footprint model (e.g. Schuepp et al. 1990, Hsieh et al. 2000), which can estimate the contribution of the total flux contribution from each upwind pixel, can be applied to the imagery and used to compute a flux aggregated from the component land cover types (Ogunjemiyo et al. 2003). This was the methodology applied in many studies for comparison of surface observations and satellite estimates (Brunsell and Gillies 2003b). Other approaches involve the aggregation of the remotely sensed vegetation and moisture characteristics (Anderson et al. 2004) and relate the spatial probability density function of surface fields to the spatial variability associated with the fluxes (Brunsell and Gillies 2003a, Kustas et al. 2004). At coarser resolutions, e.g. Moderate Resolution Imaging Spectroradiometer (MODIS) or Advanced Very High Resolution Radiometer (AVHRR), it is more common to simply conduct a direct comparison between the pixel and the observation. Given the order of magnitude of errors associated with the eddy covariance (≈10%), and the errors associated with the various modelling schemes to compute the surface flux from the satellite (French et al. 2005), agreement on the order of 10% is often considered acceptable.

It is important to note, however, that with the above approaches the atmospheric component of the transport is often only minimally considered. Therefore any impact of the footprint on the comparison between the surface observation and the satellite derived flux is an aggregation problem regarding the component surface fluxes. In reality, the surface heterogeneity may induce spatial heterogeneity within the boundary layer (Courault et al. 2009). This induces a number of potential errors when comparing surface estimates with satellite derived fluxes, as was recently discussed by Bertoldi et al. (2008). This is an issue that is only recently being considered through the combination of remotely sensed surface conditions and large-eddy simulation modelling.
Fluxes in heterogeneous terrain

Recently, the use of Large Aperture Scintillometry (LAS) has proven to be a robust, cost effective method for assessing surface energy fluxes in a variety of terrain (de Bruin 2002), offering numerous benefits over the more traditional eddy covariance technique, the primary being the ability to calculate areal averaged sensible heat fluxes over spatial distances comparable to those observed by satellites (Hemakumara et al. 2003, Lagouarde et al. 2002, Watts et al. 2000). In addition, the LAS can be used to calculate fluxes on shorter time-scales (10 minutes) (Thiermann and Grassel 1992) than eddy covariance stations which are on the order of 30 minutes or longer (Finnigan et al. 2003). This may be an important issue when considering a comparison with an instantaneous flux derived from a satellite.

The LAS technique has been applied with good results to homogeneous surfaces (Cain et al. 2001, de Bruin et al. 1995, Hoedjes et al. 2002, McAneney et al. 1995), as well as more heterogeneous environments (Lagouarde et al. 2000, Meijninger et al. 2002). Scintillometers are starting to be employed during large-scale multi-disciplinary field experiments as well; such as the SALSA campaign in the southwestern United States and northwestern Mexico (Chehbouni et al. 1999, 2000) and the LITFASS-98 experiment in northeastern Germany (Beyrich et al. 2002b). In addition, the methodology has been applied to conduct continuous monitoring over longer time periods (Beyrich et al. 2002a). While the methodology does not remove the issue of comparing the flux footprint to the pixel resolution, since the footprint of the LAS can be nearly the same size as the pixel, the issue may (or may not) be minimized (e.g. Meijninger et al. 2002, Timmermans et al. 2009). The techniques applicable to the calculation of the eddy covariance footprints are also applicable with the LAS; for instance Hoedjes et al. (2007) assessed the ability to use high resolution thermal imagery as a way to account for heterogeneity within the footprint of a scintillometer and a flux tower.

One limitation of the LAS, however, is that it requires Monin–Obukhov similarity theory to be valid. As a surface becomes more heterogeneous, this may not be the case due to variation in local land cover (Courault et al. 2009). In fact, Timmermans et al. (2009) found that in a generally unstable atmosphere, irrigation induced a local area of stable air within the footprint of the LAS. This resulted in sensible heat fluxes of opposite directions in various portions of the footprint, which the LAS is unable to detect. This potentially complicates issues when attempting to aggregate over different land cover types within a heterogeneous area (Ezzahar and Chehbouni 2009).

Ultimately, this is a scale issue involving how the heterogeneous surface interacts with the atmospheric boundary layer (Brunsell and Gillies 2003b, Anderson et al. 2003, Wu and Li 2009). The validation of remotely sensed energy and mass fluxes requires some way of scaling the surface observations to the appropriate resolution and insuring that the fluxes are originating from the same area. While the flux footprint in a remotely sensed approach is often assumed to match the pixel, in reality this may not be the case. While eddy covariance stations are the most common surface observations and are routinely used to verify remotely sensed fluxes, the towers are often located in ‘ideal’ sites which may not be representative of the true area averaged flux (Schmid and Lloye 1999). Therefore, it is essential to understand how an eddy covariance flux matches a scintillometer flux and a flux derived directly at the scale of a satellite pixel. The objective of this study was to attempt to address this issue by examining how variations in local meteorology and the flux footprint of the LAS impacts comparison with the MODIS estimates of the sensible heat flux in tall-grass prairie.
2. Site description and field measurements

Research was conducted on the Konza Prairie Biological Station (KPBS) located south of Manhattan, Kansas. This site offers approximately 3500 hectares of pristine tall-grass prairie designated for long-term research. The vegetation is tall-grass prairie, dominated by \( C_4 \) grasses. The 30-year average annual precipitation is 840 mm, with 520 mm occurring during the growing season. The site is an annually burned with silty-clay loam soils (Bremer and Ham 1999).

The tall-grass prairie is a convoluted landscape with large uplands dissected by eroded valleys and lowlands (figure 1). While the land cover class does not change across the watershed, the amount of biomass (and therefore flux) can vary significantly due to topographic position and other microclimatic variability. On average the upland aboveground biomass is 581 gm\(^{-2}\), and the lowland is 384 gm\(^{-2}\) (Heisler and Knapp 2008). In order to assess the impact of topographic variability and the associated species distribution on the spatial variability of turbulent fluxes, an eddy covariance tower is located in both an upland (441 m) and a lowland (427 m) position separated by approximately 350 metres.

![Figure 1](image_url)

Figure 1. (Top) Topographic variability of the site with path of the LAS (blue line) and the upland (U) and lowland (L) eddy covariance towers. (Bottom) Topographic variability along LAS transect.
Net carbon exchange and water vapor fluxes were measured by eddy covariance using a triaxial sonic anemometer (CSAT-3, Campbell Scientific Inc., Logan, UT, USA) and an open path gas analyzer (LI-7500, Li-Cor, Lincoln, NE, USA). The eddy flux equipment was positioned 3 m above the surface and sampled using a Campbell Scientific 23X datalogger. Wind velocities, gas concentrations, and virtual temperatures sampled at 20 Hz. The open-path gas analyzer was calibrated in the field every two weeks using tank gas and a dew point generator (LI-610, Li-Cor). Net radiation was measured with a four-component radiometer (model CNR1, Kipp and Zonen, The Netherlands) positioned 3 m above the canopy. Soil heat flux plates (Radiation Energy Balance Systems) were installed at 5 cm and the rate of change in storage above the plates was measured automatically using dual probe heat capacity sensors. Ancillary instrumentation included: a tipping bucket rain gauge, relative humidity and temperature probe (Vaisala, Helsinki, Finland), and a pyranometer and a photosynthetically active radiation (PAR) sensor (Li-Cor).

Post processing of the eddy covariance data included coordinate rotation using the planer fit method (Wilczak et al. 2001) as well as the standard corrections for density (Webb et al. 1980) and sonic-anemometer derived estimates of temperature (Schotanus et al. 1983). The protocol for the correction scheme follows that of Baum et al. (2008). Using the EdiRe software (version 1.4.3.1167, R. Clement, University of Edinburgh, UK), corrections were made for despiking, lag removal, corrections for sonic temperature heat flux corrections for humidity, sensor separation and spectral attenuation.

In addition, a Kipp and Zonen LAS was located spanning watersheds 1D and 2D. The path length of the LAS was 990 meters, and an average height along the path is 7.9 meters (determined from LIDAR derived topography). LAS data were collected at 1 Hz, and were used to compute fluxes at 10 minute time intervals. Ancillary meteorological measurements for the LAS were collected on an additional weather station. This station provided windspeeds at two heights (for computation of the friction velocity), etc. Therefore, the LAS and the eddy covariance measurements are independent of one another.

Measurements were collected for a 53 day period (DOY 198–250) in 2007. LAS data were stored at 1 Hz and $C_n^2$ values were calculated for 30 minute periods to compute the fluxes in order to compare with the eddy covariance data. For the purposes of this study, fluxes from both towers and the LAS are averaged from 10 AM to 2 PM from DOY 198 to 250, 2007.

3. Methods

3.1 Scintillometer derived fluxes

The large aperture scintillometer (LAS) measures variations in the refractive index of air by sending out a beam of incoherent radiation between a transmitter and a receiver. The scintillometer measures the structure function of the refractive index (Hill 1992) which is a measure of the atmospheric turbulence:

$$C_n^2 = \frac{n(r_1)^2 - n(r_2)^2}{r_{12}^{2/3}}$$  

(1)
where $C_n^2$ is the structure parameter for refractive index, and $n$ is the refractive index at locations $r_1$ and $r_2$ with a distance between the two points being $r_{12}$. This distance must be within the inertial subrange (Hill 1992).

Operationally, the LAS measures the log of the $C_n^2$ signal ($U_{Cn2}$), and the structure parameter for refractive index fluctuations are determined from the LAS output as:

$$C_n^2 = 10^{U_{Cn2} - 12 + 1.15\sigma_{U_{Cn2}}^2}$$  \hspace{1cm} (2)

where $\sigma_{U_{Cn2}}^2$ is the variance of the LAS output over the averaging period (e.g. 10 minutes).

Variations in the structure function are due to variations in atmospheric pressure ($p$, [Pa]), temperature ($T$, [K]) and humidity ($Q$, [g kg$^{-1}$]). The structure parameter for refractive index can then be written as a function of the structure parameters of $T$ and $Q$ (de Bruin et al. 1995):

$$C_n^2 = C_T^2 C_T^2 + 2 A_T A_Q C_T Q + A_Q^2 C_Q^2$$  \hspace{1cm} (3)

where $C$ are the associated structure parameters for $n$, $T$ and $Q$, and $C_{TQ}$ is the covariant structure parameter for $T$ and $Q$. The $A$ coefficients in equation (3) are functions of $T$, $Q$ and $p$ as well as a weak function of wavelength ($\lambda$):

$$A_T = -0.78 \times 10^{-6} \frac{p}{T} + 0.126 \times 10^{-6} R_v Q$$

$$A_Q = -0.126 \times 10^{-6} R_v Q$$  \hspace{1cm} (4)

where $R_v$ is the gas constant for vapor.

The scintillometer used in this study operates in the near-infrared ($\lambda = 880$ nm) where the contribution of the $C_Q^2$ are insignificant (Green et al. 1994). In this region, the fluctuations between the refractive index and the temperature are related by:

$$C_n^2 = \left( -0.78 \times 10^{-6} \frac{p}{T^2} \right)^2 C_T^2 \left( 1 + \frac{0.03}{\beta} \right)^2$$  \hspace{1cm} (5)

where $\beta$ is the Bowen ratio, the ratio of sensible heat ($H$, [W m$^{-2}$]) to latent heat ($LE$, [W m$^{-2}$]).

Under conditions where Monin–Obukhov similarity theory is valid, the structure parameter for temperature can be scaled to a non-dimensional form that is a function of $z/L$ where $z$ is the height [m] and $L$ is the Obukhov length [m] (Green et al. 1994). This relationship is:

$$\frac{C_T^2 (z - d)^{2/3}}{\theta_s^2} = f_T \left( \frac{z - d}{L} \right)$$  \hspace{1cm} (6)

where $d$ is the displacement height [m], and $\theta_s$ is the surface layer temperature scale [K]. The right-hand side of equation (6) has to be empirically determined as a function
of atmospheric stability. In this case we use the standard value proposed by de Bruin et al. (1995):

\[ f_T \left( \frac{z - d}{L} \right) = C_{T1} \left( 1 - C_{T2} \frac{z - d}{L} \right)^{-2/3} \]  

(7)

where \( C_{T1} = 4.9 \) and \( C_{T2} = 9 \) for unstable conditions and \( f_T (\frac{z - d}{T}) = 5 \) for stable conditions.

According to Monin–Obukhov similarity theory, the surface layer temperature scale \( \theta_s \) is given by:

\[ \theta_s = \frac{H}{\rho c_p u_*} \]  

(8)

where \( u_* \) is the friction velocity \([\text{ms}^{-1}]\), \( \rho \) is the air density \([\text{kg m}^{-3}]\), \( c_p \) is the specific heat capacity \([\text{J kg}^{-1} \text{K}^{-1}]\) and the Obukhov length is given by:

\[ L = -\frac{u_*^2 \rho c_p T}{g \kappa H} \]  

(9)

where \( g \) is the acceleration due to gravity, and \( k \) is the von-Karman constant (0.4).

Given an independent measurement of the friction velocity from either an eddy covariance station or a vertical profile of horizontal wind speeds, equations (6)–(9) can be solved iteratively to give a measurement of the sensible heat flux.

Although not of direct interest to this paper, if additional measurements of the net radiation \( (R_n, \text{[W m}^{-2}]\)) and soil heat flux \( (G, \text{[W m}^{-2}]\)) are available, the latent heat \( (LE, \text{[W m}^{-2}]\)) can be solved for by the residual of the surface energy balance:

\[ LE = R_n - G - H. \]  

(10)

### 3.2 LAS footprints

The LAS allows for computation of an areal averaged sensible heat flux. This flux originates from an upwind area, the so-called footprint, which can be approximated as a function of the mean wind speed \( (U) \) and the friction velocity \( (u_*) \) from the meteorological tower. For the purposes here we adopt the one-dimensional footprint model of Schuepp et al. (1990) and adapt it for two-dimensional variation by incorporating the weighting function of the LAS:

\[ f(x, y) = \frac{W_{LAS}(y) U_{z_{LAS}}}{\kappa u_* x^2} \exp \left( -\frac{U_{z_{LAS}}}{\kappa u_* x} \right) \]  

(11)

where \( W_{LAS} \) is the weighting function of the LAS along the path \( (y) \), \( U \) is the mean horizontal wind speed, \( z_{LAS} \) is the height of the LAS above the ground, and \( x \) is the upwind distance. Therefore, for a given deployment of the LAS, the source area contributing to the measurement is only a function of \( U/u_* \).

In order to examine the range of variability in the footprint and the relation to eddy covariance and fluxes derived from MODIS, we computed the \( U/u_* \) values for each day and have chosen to focus on the maximum and minimum values in order to
ascertain the widest range of effects on footprint distance to compare with the MODIS fluxes. Therefore, we examine two days for computation of the footprints and remotely sensed fluxes: DOY 207 ($U/u^* = 13$) and DOY 242 ($U/u^* = 6.3$).

3.3 Sensible heat flux derived from MODIS

Spatially distributed fluxes are determined from the MODIS fractional vegetation ($Fr$) and radiometric temperature ($Tr$) using the ‘triangle’ method (Gillies et al. 1997, Brunsell and Gillies 2003b, Carlson 2007). This method relies on the assumption that an image contains the entire range of vegetation cover and near surface soil moisture ($M_0$). Assuming these values are present, the model relies on coupling a soil-vegetation-atmosphere-transfer (SVAT) model with an atmospheric boundary layer (ABL) model to compute the fluxes. Here, the Simsphere SVAT model (Carlson et al. 1995) was used to determine the fluxes from the MODIS imagery. The model was initialized with a radiosonde profile and the general site characteristics present at the Konza Prairie. The model description is covered in detail elsewhere (e.g. Taconet et al. 1986, Lynn and Carlson 1990), and recently Brunsell et al. (2008) reviewed the methodology for derivation of the latent heat flux.

Prior to calculation of the fractional vegetation, the MODIS imagery must be corrected for atmospheric effects. Here, the daily MODIS scenes were corrected using the methodology of Brunsell and Gillies (2002). This methodology uses the MODTRAN radiative transfer model to correct the visible, near-infrared and thermal infrared bands. A radiosonde profile is required for each day of analysis to accurately characterize the radiative properties of the atmosphere. The radiosonde profiles used in this analysis were obtained from the Topeka, KS National Weather Service office approximately 50 miles east of the Konza Prairie. MODTRAN is iterated over scan-angle and (in the case of the thermal infrared correction) surface emissivity values (Brunsell and Gillies 2002). Thus, following the correction algorithm, the at-sensor temperature has been corrected to a surface radiometric temperature value.

The fractional vegetation values are computed from a scaled version of the Normalized Difference Vegetation Index (NDVI) corresponding to the time of satellite overpass on each day of analysis. The NDVI is calculated from the atmospherically corrected surface reflectance in the red ($\rho_r$) and near-infrared ($\rho_n$) bands:

$$\text{NDVI} = \frac{\rho_n - \rho_r}{\rho_n + \rho_r}$$

from which the pixel scale fractional vegetation can be calculated as (Gillies and Carlson 1995):

$$Fr = \left( \frac{\text{NDVI}_0 - \text{NDVI}_v}{\text{NDVI}_v - \text{NDVI}_s} \right)^2$$

where NDVI$_0$ is the pixel NDVI value, and the $v$ and $s$ subscripts denote the vegetation and bare soil values respectively. These values are approximated from visual inspection of the imagery. In general, the vegetated value corresponds to the maximum NDVI value, while the soil value corresponds to the minimum non-water or cloud pixel.
For computing the fluxes from the remotely sensed data, model calibration must be done for each day of interest. The calibration is conducted by fitting the model to the four corners corresponding to saturated bare soil \([Fr = 0, M_0 = 1]\), dry bare soil \([Fr = 0, M_0 = 0]\), saturated vegetation \([Fr = 1, M_0 = 1]\) and dry vegetation \([Fr = 1, M_0 = 0]\). Following the calibration, the SVAT is iterated over combinations of soil moisture availability in 20% increments and fractional vegetation in 20% increments. The model predicted fluxes and output radiometric temperature is then used to construct a polynomial regression according to:

\[
F = \sum_{j=0}^{3} \sum_{i=0}^{3} a_{ij} Fr^i Tr^j
\]  

where \(F\) is the modelled flux of interest (e.g. \(H\), \(LE\), etc.), \(Fr\) is the input fractional vegetation, \(Tr\) is the model output radiometric temperature, and the \(a_{ij}\) are the regression coefficients.

The derived regression equations are then applied to the remotely sensed \(Fr\) and \(Tr\) values to determine the fluxes. As stated above, for the purposes of comparison with the widest range of effects, we only computed the MODIS fluxes for two days within the study time frame (DOY 207 and 242). Also note that the \(Tr\) values are corrected land surface temperatures (i.e. not brightness temperatures). Of course, any errors in the remotely sensed temperature will translate to errors in the derived flux values and it is important to have the highest quality surface temperature dataset.

4. Results

4.1 Temporal variability of LAS measurements

The sensible heat fluxes derived from the LAS are shown in figure 2. A small gap is present due to sensor malfunction with the ancillary measurements and therefore a
heat flux independent of the eddy covariance towers could not be obtained. Nighttime values from a scintillometer are often problematic due to the inability to determine the correct direction of the flux, but in this case most values seem reasonable (small negative values).

Since our primary focus is the comparison with the remotely sensed values, we average the scintillometer fluxes during midday hours (10 AM to 2 PM LST) to obtain midday sensible heat values for comparison with both the eddy covariance towers and the MODIS fluxes. This raises a number of issues including the assumption that the fluxes derived from the time of satellite overpass (≈11 AM) are comparable to an average midday observation. Given that the MODIS data used here was obtained on cloud-free days, this is probably not an unrealistic assumption. However, it is possible that errors are introduced due to the representativeness of the time of the satellite overpass to the average midday fluxes.

Next, we present the two days of scintillometer measurements that correspond to the two days of special interest DOY 207 and 242 (figure 3). The structure parameter for the refractive index for both days is shown in the bottom panels for the midday times of 10 AM to 2 PM local time. DOY 207 corresponds to the larger footprint \((U/u^*) = 13\) and the \(C_n^2\) fluctuations are generally smaller than for DOY 242 \((U/u^*) = 6.3\).

A wavelet analysis of the \(C_n^2\) data was conducted in order to examine the temporal scales contributing to the LAS sensible heat flux. The top panels of figure 3 show an example of the half-plane (red higher energy, green/blue lower energy) with the corresponding time series of the \(C_n^2\) data. The wavelet spectra are computed for the two days of special interest and plotted in figure 4. The wavelet spectra demonstrate a significant contribution from time-scales larger than 30 minutes on both days. This could potentially point to problems with using the half-hour time periods for computation of both the LAS and eddy covariance fluxes.

### 4.2 Comparison of LAS and eddy covariance fluxes

As a first step to comparison with the remotely sensed fluxes, we compare the LAS with the two eddy covariance towers in the watershed. The LAS and eddy covariance midday averaged daily sensible heat fluxes for DOY 198–250, 2007 are shown in figure 5. In general, agreement is much better between the upland tower and the LAS, which is illustrated by a linear regression of \(H_{LAS} = 0.94H_{EC,UP}\). The overall correlation is 0.96, with a root mean squared error (RMSE) of 7.4 Wm\(^{-2}\) and a bias of \(-6.2\) Wm\(^{-2}\). The lowland site shows slightly poorer agreement with the LAS: \(H_{LAS} = 0.86H_{EC,LOW} + 25.9\), with a correlation of 0.84, RMSE of 13.6 Wm\(^{-2}\), and a bias of 17.7 Wm\(^{-2}\).

Next, we examine the difference between the towers and the scintillometer as a function of various wind conditions (figure 6). The upland tower is typically a higher value than the LAS, resulting in a negative \(\Delta H\). This is expected since the upland regions are associated with shallower, drier soils and less biomass. This would lead to higher \(H\) values than might be representative of the areal average. Likewise, the lowland tower shows the opposite when compared to the LAS, due to higher biomass and moisture conditions in the lowlands leading to a higher latent heat flux and a reduced sensible heat flux. Since the LAS footprint covers a variety of upland and lowland conditions and a larger footprint than either of the towers, the LAS measurements are generally between the two extremes and therefore are more representative of the areal averaged flux.
Figure 3. Wavelet decomposition of the $C_n^2$ LAS signal and the time series of the $C_n^2$ for (top) DOY 207 and (bottom) DOY 242.

Figure 6 illustrates the differences ($\Delta H = H_{LAS} - H_{EC}$) between the LAS and the towers as a function of wind direction. The majority of the winds are out of the S-SW, with little real relationship with the magnitude of the difference between the
Figure 4. Wavelet spectra for DOY 207 and DOY 242.

Figure 5. Comparison of sensible heat fluxes measured by the upland and lowland eddy covariance towers and the scintillometer.
Fluxes in heterogeneous terrain

Figure 6. Difference between LAS and eddy covariance measurements of $H$ as a function of mean windspeed, wind direction, friction velocity and ratio of $U/u^*$. scintillometer and the towers. There is also little correlation between the flux differences and the mean wind speed $\bar{U}$. There may be a slight relationship with the friction velocity as the values decrease with increasing $u_*$. In addition, note the possible weak relationship with the upland tower that as the footprint ($U/u_*$) increases the difference between measurements seems to decrease, although this is not observed at all with the lowland tower.

Therefore we conclude that while the overall agreement between the scintillometer and the towers is good, the variation is not due to systematic variation in the wind conditions.

4.3 Sensible heat fluxes from MODIS

Next we calculated the sensible heat flux from the MODIS sensor for the two days of interest. The associated joint probability density functions between the radiometric temperature and fractional vegetation are shown in figure 7. These are the ‘triangle’ plots to which the SVAT model is calibrated in order to determine the fluxes. The overall temperature variation for a majority of the pixels is on the order of $6^\circ C$ for
both days; the shape of the triangle does exhibit significant variability between the two days.

The MODIS fluxes determined for days 207 and 242 are shown in figure 8. These maps illustrate a larger geographic extent than that presented in figure 1 due to the assumptions of the moisture and vegetation variability. Note that while day 207 actually has the larger surface temperatures, the resultant heat fluxes are smaller than those from DOY 242. This simply reaffirms that it is impossible to use solely the radiometric temperature to determine the sensible heat flux.

4.4 LAS footprint and MODIS fluxes

Before we compare the scintillometer measurements to the MODIS derived fluxes, we must compute the footprint for the LAS. The LAS footprint for the two values of $U/\mu_*$ are shown in figure 9 assuming a wind from the south. The colored area shows the topographic variability contributing to 90% of the observed flux. It is obvious that the area contributing to the flux is larger for DOY 242.

By convolving the LAS footprint (after rotation to the mean wind direction, 255° on DOY 207 and 235° on DOY 242) with the MODIS derived sensible heat fluxes we
Figure 9. Variation in source area contributions for (left) DOY 207 and (right) DOY 242.

are able to compare the remote sensing values with the LAS (figure 10). Table 1 shows the values of each of the towers, LAS and remotely derived fluxes for the two days of interest. Note that the MODIS derived \( H \) value corresponds better to the LAS measurements on DOY 242 (percentage error: 0.04), which contained the larger footprint \( (U/u_\ast = 6) \), while on DOY 207 \( (U/u_\ast = 13) \) the MODIS value is significantly lower but with a larger percentage error \( (-0.13) \).

There is some measurement and modelling error associated with each of the values in table 1. While it is outside the scope of this paper, it is important to note that there may in fact be little difference between each of these estimates. This is, in fact, why many estimates of fluxes using remote sensing data are often deemed ‘acceptable’.

5. Conclusions

We investigated the variability in the surface contribution to a measured sensible heat flux. Using a relatively coarse flux determined from a satellite, we have shown that the upwind contributing area can be quite significant on the derived flux. Comparison with eddy covariance towers and scintillometry illustrate the location bias often observed in tower locations, and points out the necessity for considering the nature of the spatial aggregation of the flux prior to comparison with satellite estimates.

In addition, the results of this study imply that:

1. The difference between the LAS and eddy covariance towers may also be a function of the \( U/u_\ast \) for the upland tower. This implies that the larger the footprint contributing to the LAS (and the eddy covariance towers) the better the agreement. Further research on this topic is warranted.

2. The MODIS derived fluxes agree to some extent with the surface measurements, but the smaller the footprint the worse the agreement (for the two days used in this study). This combined with energy balance issues and time-scales of atmospheric processes in the boundary layer may limit the routine application of remote sensing for the determination of spatially variable surface energy fluxes as well as comparison with the more numerous eddy covariance flux towers.
Figure 10. MODIS derived sensible heat flux as would be measured by the LAS for (top) DOY 207 and (bottom) DOY 242.

Table 1. Comparison of sensible heat fluxes (W m\(^{-2}\)) from the eddy covariance towers, LAS and MODIS.

<table>
<thead>
<tr>
<th>DOY</th>
<th>EC (lowland)</th>
<th>EC (upland)</th>
<th>LAS</th>
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Fluxes in heterogeneous terrain


