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Heat storage and energy balance fluxes for a temperate deciduous forest

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Abstract

Hourly observations of forest-atmosphere energy balance components are presented for the Morgan-Monroe State Forest, south-central Indiana, USA for the period March 1998 to December 2001, with particular emphasis on the storage heat flux (ΔQ_S). The sub-components of ΔQ_S showed strong diurnal variability, although ground heat flux dominated the seasonal change of storage heat flux. The annual storage heat balance for the three complete years showed a small but consistent deficit averaging 16.18 MJ m^{-2} a^{-1}. On a per-hour basis this is small compared with potential error highlighted in sensitivity testing of a range of methodological issues. Under-sampling of biomass temperature yielded the greatest uncertainty. Energy balance closure, determined by linear regression, found turbulent fluxes to underestimate available energy by 28% ($r^2 = 0.92$), with no statistically significant difference in closure found between years or between months. On a diurnal basis, weaker closure was associated with low friction velocity and low turbulent heat flux magnitude. Increasing turbulent flux block averaging time from 15 to 60 min resulted in a small improvement in closure at 60 min averages (2.5%). In addition, co-spectral corrections to the latent heat flux to account for long-tube damping effects generated a much larger improvement in closure (16%). The mean annual Bowen ratio was 0.59, ranged from 0.57 to 0.61 and inter-annual variability of all energy balance components over the period was very small (less than 6% departure from mean).

Keywords: Energy balance; Forest-atmosphere flux; Forest heat storage; Energy balance closure.

1. Introduction

Ecosystem–atmosphere exchanges of energy display considerable variability over a range of temporal and spatial scales. The pattern of variability is a

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function of complex interaction between atmospheric and biophysical controls. Multi-year datasets of hourly energy balance observations provide a useful resource for evaluating the behavior of surface–atmosphere interaction across timescales with five orders of magnitude. In a forest environment, the surface energy balance is conveniently expressed in terms of the energy transfer into and out of a control volume including the vegetation, the canopy-air space and a layer of soil deep enough to exclude heat transfer through its bottom, following (Oke, 1987):

$$Q^{*} = Q_{\mathrm{H}} + Q_{\mathrm{E}} + \Delta Q_{\mathrm{S}} + \Delta Q_{\mathrm{A}} \tag{1}$$

where Q^{*} is net all-wave radiation, Q_{H} is sensible heat flux, Q_{E} is latent heat flux (all through the top of the volume), ΔQ_{S} is the storage heat flux within the volume and ΔQ_{A} is the advective heat flux through all sides of the volume. Numerous difficulties and assumptions in measurement of energy balance components over and within tall vegetation contribute to the lack of conservation of energy observed at many sites (e.g. Lee and Black, 1993; Gay et al., 1996; Baldocchi and Vogel, 1996; Blanken et al., 1997, 1998; Wilson and Baldocchi, 2000; Wilson et al., 2002). Of the components in (1), ΔQ_{S} and ΔQ_{A} are the most difficult to determine and have often been neglected over time periods significantly longer than a day. Here, we focus attention on the estimation of ΔQ_{S} by examining sub-components of heat storage within the forest, soil and air column and discuss strategies to scale up to an area that is comparable to the expected footprints of turbulent fluxes and radiation. However, because the forest ecosystem of our site is horizontally extensive (see Section 2), we assume that advection of heat is negligible.

This paper reports 4 years of observations of surface energy balance components (except ΔQ_{A}) over a mid-latitude, mixed deciduous forest. Objectives of the research are threefold. The first is to evaluate the methods for estimating surface energy balance components, particularly the storage heat flux. Past research reveals a range of methods with varying degrees of complexity employed to determine ΔQ_{S} (e.g. McCaughey, 1985; McCaughey and Saxton, 1988; Jaeger and Kessler, 1996; Meesters and Vugts, 1996; Vogt et al., 1996; Blanken et al., 1997; Ohta et al., 1999; Wilson and Baldocchi, 2000). Given the variability of methods and emergence of multi-site comparisons (Wilson et al., 2002; Curtis et al., 2003), emphasis is placed on determining the sensitivity of ΔQ_{S} to a range of flux components and parameters. The second objective is to evaluate the quality of energy balance closure under a range of conditions in order to identify possible causes for lack of closure in tall vegetation. This is compared with a recent study of energy balance closure at multiple terrestrial ecosystem flux tower sites by Wilson et al. (2002) and assesses the energy balance implications of co-spectral corrections of Su et al. (2004). The third objective is to identify the dominant controls of temporal variability of energy balance components from diurnal to interannual. Comparisons are also made with multi-year energy balances reported for a mixed deciduous forest at Oak Ridge, Tennessee (Wilson and Baldocchi, 2000) to place the current analysis in a regional context.

2. Site and methods

Observations are made from and around a 46 m tower located in the Morgan-Monroe State Forest (MMSF) in south-central Indiana, USA (39°19′N, 86°25′W). The MMSF flux tower is used for observations of CO_2, heat, water vapor and radiation fluxes as part of the AmeriFlux and FLUXNET observational networks. MMSF is a managed forest with a total area of 95.3 km^2 (Schmid et al., 2000). The area has a ridge-ravine topography with a relative relief of <60 m, and an overall elevation range of <100 m. Soils are well drained inceptisols, formed in residuum from sandstone, siltstone, and shale (USDA, 1980). MMSF is a successional broadleaf forest within the maple-beech to oak-hickory transition zone of the eastern deciduous forest, with a mean canopy height near the tower of 27 m. Twenty-nine tree species have been identified in the area surrounding the tower. Total basal area (sum of cross-sectional area of tree boles at 1.39 m height) is 26 m^2 ha^{-1}. Five tree species comprise 73% of the total basal area: sugar maple (*Acer saccharum*, 27%), tulip poplar (*Liriodendron tulipifera*, 19%), sassafras (*Sassafras albidum*, 9.5%), white oak (*Quercus alba*, 9%), and black oak (*Quercus nigra*, 8.5%). Ages of representative trees were determined from increment cores; many of the trees in the immediate area of the tower are about 55 years old. The mature trees (dbh >7 cm) in a 78.5 ha area (defined by a 500 m radius around the tower) have been inventoried using a series of small plots along radial transects. Although some of the forest stands in this area experienced light selective timber harvesting in 1974 and a small area (ca. 3–4 ha) experienced severe wind damage in 1990, the forest surrounding the tower is fairly uniform in tree species composition, size and age. Based upon measurements in 54 10 m × 15 m plots and using allometric equations, total aboveground biomass was estimated to be 19.52 kg m^{-2} (Ehman et al., 2002). Seasonal phenological observations from 1998 to 2001 show leaf emergence occurring on average on 9 April with average senescence on 30 October.

The MMSF flux tower supports eddy covariance, radiation and meteorological instrumentation at multiple levels. The minimum fetch of uninterrupted forest exceeds 4 km in any direction and reaches up to 8 km in the principal wind direction (west to southwest). Four micrometeorological stations located within the expected footprint near the forest floor, provide additional spatial information on below canopy conditions. Schmid et al. (2000) provide an overview of instruments and sampling characteristics in the MMSF. The following sections outline the methods used to estimate the terms of the energy balance (1) and discuss sources of uncertainty. Instrumentation at MMSF is not currently configured to measure ΔQ_A, but its potential role is discussed in Section 3.2.

2.1. Net radiation

Net radiation observations are made at 46 m using a Kipp & Zonen CNR1 net radiometer (Kipp & Zonen Corp., Delft, Holland). Vogt et al. (1996) discussed sources for error in net radiation measurement, including calibration drift over time, inaccurate leveling of instruments, variability of dome transmissivity, sensor response with wavelength and unequal losses of energy through convection and thermal infrared emission between upward and downward pointing sensors during periods of high irradiance intensity. However, instrument intercomparison studies typically reported small magnitudes of error; for example, 2% (Laubach and Teichmann (1999), 3.5% (Halldin and Lindroth, 1992; Vogt et al., 1996), and a maximum absolute error of 30 W m^{-2} (Oncley et al., 2002). In this study, radiometers are leveled and domes cleaned at least monthly. A second CNR1, with a new factory calibration, was used as a reference at 46 m over two 2-week periods, (summer 1998, summer 2001). In both cases, the permanent radiometer was found to overestimate 15 min average net radiation fluxes of the intercomparison radiometer by less than 2.5% ($R^2 > 0.99$).

2.2. Sensible and latent heat fluxes

The main instruments used for the eddy covariance system are a CSAT-3 three-dimensional sonic anemometer thermometer (Campbell Scientific Inc., Logan, UT) and a LI-6262 closed path gas analyzer (LiCor, Lincoln, Nebraska). The CSAT-3 is situated at 46 m and air is drawn from an intake adjacent to the sonic head through sample tubes to the gas analyzer located in a small building at the base of the tower. Measurements of water vapor concentration are made by the gas analyzer in differential mode using N_2 as a reference gas. Measurements of three-dimensional turbulent velocity (m s^{-1}) and virtual temperature (K) from the CSAT-3 and water vapor mixing ratio (from the LI-6262, determined as the mole fraction to the dry air reference, to account for the WPL effect, Webb et al., 1980) are sampled at a frequency of 10 Hz. Vapor eddy covariance fluxes are calculated by shifting the time series of water vapor mixing ratio (which contains a time lag of approximately 7 s due to transport down the sample tube) and vertical velocity to maximize the correlation coefficient between them. Covariances between vertical velocity fluctuations and either virtual temperature or water vapor are calculated over 15 min periods. Spikes in all components are determined either by instrument or post-processing diagnostic flags as described in Schmid et al. (2000). In a second phase of quality control, all measured or derived variables (15 min averages) are submitted to statically defined physical plausibility constraints. Subsequently, the 15 min data are reduced to hourly averages.

Intercomparison of Q_E was made over a 2-day period between the described measurement and processing system and an AmeriFlux roving system and very close agreement for Q_E was found between the two systems, with a regression slope of 1.003 and a coefficient of determination (R^2) of 0.99 (Schmid et al., 2000; AmerFlux Roving System, 2003). Differences between the two systems included the block averaging periods (15 min in this study versus 30 min in the roving system). Blanken et al. (1997) also found overall differences of less than 10% between 15 and 30 min block averaging periods over a boreal aspen forest. Co-spectral corrections to long-tube damping effects in the use of closed path gas analyzers were shown to have an increasing impact on latent heat fluxes over time at two forest sites (MMSF and the site at the University of Michigan Biological Station in northern lower Michigan) by Su et al. (2004). The influence of the eddy-covariance averaging period and co-spectral corrections on energy balance closure is assessed in Section 3.2.

2.3. Storage heat flux

Unlike the three components discussed above, ΔQ_S is not measured directly but estimated from a number of constituent components. Methods, definitions and assumptions used to determine ΔQ_S have varied widely in the past but McCaughey and Saxton (1988) defined ΔQ_S comprehensively as the sum of contributions from several ecosystem compartments:

$$\Delta Q_S = \Delta Q_g + \Delta Q_a + \Delta Q_w + \Delta Q_v + \Delta Q_p \qquad (2)$$

All components in (2) are expressed as heat storage change fluxes (in W m^{-2}), where ΔQ_g refers to the soil, ΔQ_a and ΔQ_w are the sensible and latent heat contributions in the air-column below the eddy-covariance system, ΔQ_v is the biomass heat storage flux, and ΔQ_p is the photosynthetic heat component. Positive values indicate transfer of energy to storage. Traditionally, many studies have ignored one or more components, particularly ΔQ_p (Jaeger and Kessler, 1996; Vogt et al., 1996; McCaughey, 1985; McCaughey and Saxton, 1988; Ohta et al., 1999; Wilson and Baldocchi, 2000), and some also excluded ΔQ_a and ΔQ_w (Wilson and Baldocchi, 2000) and ΔQ_v (Ohta et al., 1999). Thom (1975); Jarvis et al. (1976) and Tajchman (1988) found ΔQ_p to be negligible, although Blanken et al. (1997) illustrated that, on a 30 min basis, ΔQ_p can amount up to 23% of ΔQ_S on clear summer days, with the contribution of foliage in ΔQ_v somewhat smaller (less than 4.6%). Because Schmid et al. (2000) indicated an overall light efficiency for photosynthetic carbon assimilation of less than 2% for this site, we follow these studies to ignore ΔQ_p, but examine the potential magnitude of leaf-ΔQ_v relative to the heat storage change in the canopy air-space, ΔQ_a. All other components of (2) are estimated based on measurements.

2.3.1. Soil heat storage flux

The soil heat storage flux (ΔQ_g) is calculated as:

$$\Delta Q_g = Q_g(z) + C_S \frac{\Delta T_S}{\Delta t} z \qquad (3)$$

where Q_g (z) is the measured soil heat flux at depth z (m), T_S is the average soil temperature (K) above the heat flux plates, t is time (in this case $\Delta t = 1$ h) and C_S is the soil heat capacity calculated from:

$$C_S = \rho_b c_{sd} + \theta_v c_{sw} \qquad (4)$$

where ρ_b is the bulk density of the soil, c_{sd} and c_{sw} are the specific heats of the dry mineral soil (c_{sd} = 890 J kg^{-1} K^{-1} for clay soils) and the soil water (c_{sw} = 4190 J kg^{-1} K^{-1}) respectively, and θ_v is the volumetric water content (%) in the soil measured using CS615 Water Content Reflectometers (CSI, Logan, UT). ΔQ_g is sampled at four locations in the forest near the main tower using pairs of heat flow transducers (REBS, Seattle, WA) inserted on average at 6.5 cm below the surface with spatial averaging type E thermocouples (TCAV, CSI, Logan, UT) to measure soil temperature at two locations in the column of soil above the heat flow transducers. A soil bulk density (ρ_b) mean value of 1.2 kg m^{-3} was used, as determined by Ehman et al. (2002) for 0–10 cm depths.

2.3.2. Canopy air-space sensible heat storage flux

The sensible heat storage flux in the canopy air-space (ΔQ_a) is calculated following, e.g., McCaughey (1985):

$$\Delta Q_a = \int_0^{z_r} \rho c_p \frac{\partial T}{\partial t} dz \cong \rho c_p \sum_{i=1}^{n} \left(\frac{\Delta T_a}{\Delta t} \Delta z_i \right) \qquad (5)$$

where ρ is the air density, c_p is the specific heat of air, z_r is the height of the net radiation measurement (46 m), T is air temperature (K) in the air-column below z_r, and T_a is a representative layer-average of T in each of several layers, Δz_i, below z_r. Thom (1975) and McCaughey (1985) argued that it is reasonable to assume ρ and c_p to be constant within the canopy layer. Thus, $\Delta T_a/\Delta T$ is often calculated as an average of measurements at several levels in the canopy (Lee and Black, 1993; Vogt et al., 1996; Blanken et al., 1997). In this study, T_a was measured at six levels (1, 6, 12, 22, 34, 46 m) using ventillated HMP35C thermistors (Vaisala, Boulder, Co). ΔQ_a was calculated by summing (5) through the six levels of temperature measurements using hourly increments. Comparison was made between this approach and linear averaging of temperature profiles or using a single temperature measurement. Linear average profiles tended to over-estimate ΔQ_a slightly due to the much faster changes occurring near the surface (mean absolute difference (MAD) = 0.86 W m^{-2}) while using a single measurement the agreement ranged from a 7% under-estimation at 46 m to a 9% over-estimation at 1 m, with the closest agreement coming at 22 m (MAD = 01.23 W m^{-2}), and the least agreement at 1 m (MAD = 4.10 W m^{-2}).

2.3.3. Canopy air-space latent heat storage flux

The latent heat storage flux (ΔQ_w) is calculated following, e.g., Thom (1975) or McCaughey (1985):

$$\Delta Q_w = \int_0^{z_r} \rho L_e \frac{\partial e_i}{\partial t} dz \cong \rho L_e \sum_{i=1}^{n} \left(\frac{\Delta e_i}{\Delta t} \Delta z_i \right) \quad (6)$$

where L_e is the latent heat of vaporization and e is vapor pressure, calculated from T_a and relative humidity measurements from the HMP35C profile. Similarly to ΔQ_a, (6) was calculated for the six levels to 46 m and summed. Comparison of ΔQ_w computed in this way to that using a single humidity sensor showed larger error than was found for ΔQ_a, ranging from the most representative height, 22 m (MAD = 4.35 W m^{-2}) to the least, 1 m (MAD = 7.49 W m^{-2}).

2.3.4. Biomass heat storage flux (boles and branches)

Biomass heat storage flux has received spotty attention over the last several decades. The theoretical treatment of this topic is mathematically rather involved and lends itself to tractable solutions only after stringent simplifications (Meesters and Vugts, 1996; Herrington, 1969). Practical applications are challenged by the sheer variety of shapes, sizes, orientations, and materials that constitute the biomass in a forest. In light of these formidable difficulties, most practical studies of biomass heat storage change have followed simplified experimental approaches proposed by Thom (1975), McCaughey (1985) or Moore and Fisch (1986). Here, we adopt the simplest of these methods (i.e., McCaughey, 1985). Our results suggest that compared to other energy balance components, the biomass heat storage flux magnitude is mostly too small to justify a more intensive effort for its estimate.

Table 1
Tree species sampled for estimation of Q_v at MMSF, including biometric data (J.C. Randolph, personal communication) and measurement height of thermocouples

	Tree species (common name)					
	Tulip Poplar	Sugar Maple	White Oak	Red Oak	White Ash	Sassafras
Tree height (m)	33.0	29.0	34.0	40.0	33.5	16.4
Tree dbh (cm)	85.0	28.6	60.5	85.9	65.9	10.2
m_{bole} (kg m^{-2})	0.551	0.045	0.275	0.644	0.339	0.004
m_{branch} (kg m^{-2})	0.112	0.011	0.059	0.129	0.071	0.001
Thermocouple placement	Height above ground (m)					
Bole	4.0	2.0	2.0	2.0	1.9	1.3
	12.0	10.0	10.0	12.0	8.0	4.4
	27.2	16.3		21.0	21.0	14.5
				30.5	26.0	
				30.5		
Branch	16.1	12.0	19.4			
	17.8	14.1	19.5			
	20.9	14.7	25.5			
	21.5	15.7	25.5			
			27.3			

According to McCaughey (1985), (see also Thom, 1975; Wilson and Baldocchi, 2000) the biomass heat storage flux, ΔQ_v, can be obtained with a simple calorimetric approach:

$$\Delta Q_v = m_{veg} c_{veg} \frac{\Delta T_{veg}}{\Delta t} \tag{7}$$

where m_{veg} is the mass of vegetation per unit horizontal area for the forest, c_{veg} is a representative specific heat of vegetation, T_{veg} is a representative biomass temperature. At MMSF, T_{veg} was estimated using 33 probes (T-type thermocouples, Omega, Stamford, CT) inserted 2 cm into the sapwood of boles and branches of six tree species at heights ranging from 1.3 to 30.5 m (Table 1). Accordingly, this method was applied separately to boles and branches to obtain the contributions of $\Delta Q_{v,\text{ bole}}$ and $\Delta Q_{v,\text{ branch}}$. A slightly different approach was followed to estimate the contribution of foliage to vegetation heat storage flux, $\Delta Q_{v,\text{ leaf}}$, and will be discussed in the next sub-section.

The specific heat of vegetation (c_{veg}) is a value that is either calculated (Moore and Fisch, 1986; Blanken et al., 1997) or assumed (McCaughey, 1985; McCaughey and Saxton, 1988; Vogt et al., 1996; Wilson and Baldocchi, 2000), and most studies apply a single value to all vegetation species considered. Moore and Fisch (1986) used a value of 2650 ± 80 J kg^{-1} K^{-1} which they estimated for Amazonian rain forest. Wilson and Baldocchi (2000) used a combined specific heat for woody biomass and water (3340 J kg^{-1} K^{-1}) for a deciduous forest in Tennessee. Thom (1975) argued that c_{veg} could be assumed to be 70% of the specific heat of water (c_w) (i.e. 2928 J kg^{-1} K^{-1}) and this value was used by McCaughey (1985) and McCaughey and Saxton (1988) for a boreal forest. In this study we also used Thom's (1975) estimate of c_{veg}. To examine the sensitivity of this parameter, we compared our results of ΔQ_v with those derived using Moore and Fisch's (1986) c_{veg} value, (resulting in MAD, of 0.45 W m^{-2}) and using Wilson and Baldocchi's (2000) value (giving a MAD of 0.66 W m^{-2}). We thus conclude that the sensitivity of the estimate of ΔQ_v to the exact value of c_{veg} is small.

Measurements of diameter at breast height (dbh, cm), tree density (individuals per m^2) and bole and branch biomass (m_{bole} and m_{branch}, kg dry mass per m^2 horizontal surface area) were obtained for the six dominant tree species where T_{veg} was measured, as well as 23 other species in smaller numbers in the 54 150 m^2 plots in the vicinity of the MMSF flux tower. The bole (m_{bole}) and branch (m_{branch}) biomass per unit horizontal area were determined for each species using allometric equations for multiple deciduous species (Harris et al., 1973).

Biomass heat storage was calculated individually for the six tree species where T_{veg} was measured, using (7). For each measured tree species, T_{veg} was averaged for the boles ($T_{v,\text{ bole}}$) and also for the branches ($T_{v,\text{ branch}}$) and then ($\Delta T_{v,\text{ bole}}/\Delta t$) and ($\Delta T_{v,\text{ branch}}/\Delta t$) were calculated, where Δt was 1 h.

The species-specific contributions of boles and branches, $\Delta Q_{v,\text{ bole}}$ and $\Delta Q_{v,\text{ branch}}$, were integrated to a species-total value, $\Delta Q_{v,\text{ species}}$. An estimate for the forest, $\Delta Q_{v,\text{ forest}}$, can be obtained by summing over all $\Delta Q_{v,\text{ species}}$ in the forest:

$$\Delta Q_{v,\text{forest}} = \sum\nolimits_{i=1}^{n_s} \Delta Q_{v,\text{species},i}, \tag{8}$$

where n_i is the number of species in the forest. However, since T_{veg} was measured only for the six dominant species at MMSF (67% of biomass), heat storage in the remaining biomass was calculating using T_{veg} from measured species with the closest dbh to each of the unmeasured species, thereby relating thermal mass with heat exchange (Aston, 1985).

A solution to (7) assumes that a representative biomass temperature, T_{veg}, can be determined (McCaughey and Saxton, 1988). We compared the hourly values of integrated $\Delta Q_{v,\text{ forest}}$ with those determined from single thermocouple measurements in a range of species and heights (Table 1), using 33 thermocouples applied independently to the whole forest m_{veg} (19.52 kg m^{-2}). In general, individual measurements scattered widely around $\Delta Q_{v,\text{ forest}}$, ranging in MAD from 2.44 to 8.82 W m^{-2}. This suggests that under-sampling of species and/or measurement heights is likely to introduce variability unrepresentative of the total biomass, particularly in diverse and tall forests. In some studies ΔQ_v has also been estimated using the ambient air temperature as a surrogate for T_{veg} (Tajchman, 1971; Thom, 1975). To evaluate this alternative, we recast our analysis of $\Delta Q_{v,\text{ forest}}$, but based on the measured air temperature profile, and applied the results to the whole forest m_{veg}, both temperatures for individual heights and the layer-weighted average temperatures. Interestingly, air temperature based Q_v produced a closer fit to the integrated $Q_{\mathrm{v,\ forest}}$ than 17 of the 33 individual thermocouples. The $\Delta Q_{\mathrm{v,\ forest}}$ based on the weighted average air temperature was closest to that based on bole and branch temperature measurements (MAD = 4.09 W m^{-2}), although the range was very small. This may be reduced somewhat by accounting for the lag between air temperature and vegetation temperature at 2 cm depth, which was found to average about 1.5 h (Zutter, 2001).

2.3.5. Foliage heat storage flux

As mentioned, the contribution of foliage to the heat storage flux ($\Delta Q_{\mathrm{v,\ leaf}}$) is commonly assumed small and neglected. For example, Blanken et al. (1997) illustrated that, on a 30 min basis, $\Delta Q_{\mathrm{v,\ leaf}}$ is commonly less than 4.6% of ΔQ_S for 30 min fluxes on clear summer days at a boreal forest site. However, to estimate the magnitude of $\Delta Q_{\mathrm{v,\ leaf}}$ at the MMSF site, several considerations and assumptions were made.

Broad leaves, such as found in MMSF, have a high surface area per volume (of leaf). The leaf temperature of transpiring leaves exposed to the sky (i.e., sunlit leaves in daytime) is usually found to be slightly less than the local air temperature (T_a) during the day, and lower than T_a at night (due to the high emissivity and sky view factor), whereas leaves with weaker radiative forcing follow air temperature closely (see, e.g., Campbell and Norman, 1998). A conservative estimate for the daily leaf temperature cycle is thus to assume equality with the local air temperature cycle. Based on these considerations, the magnitude of $\Delta Q_{\mathrm{v,\ leaf}}$ can be assessed relative to ΔQ_a, following (5), but using an appropriate value of heat capacity for foliage.

In 1999 and 2000, Ehman et al. (2002) reported a leaf detritus production of about 200 gC m^{-2} year^{-1}. Neglecting any evergreens and herbacious layer leaves not captured in the leaf-litter traps, this amount is assumed to represent the maximum carbon mass per unit ground in the foliage, $\hat{m}_{\mathrm{fol,C}}$. From this value we obtain the total mass (per unit ground) of foliage, m_fol, by the following steps: (i) Carbon constitutes about 46% of dry biomass (C. Wayson, personal communication), and thus peak dry leaf mass is $\hat{m}_{\mathrm{fol,d}} \approx \hat{m}_{\mathrm{fol,C}}/0.46$. (ii) About 70–80% of mass in tree leaves is water (D. Parkhurst, personal communication), and thus peak total leaf mass is $\hat{m}_\mathrm{fol} \approx \hat{m}_{\mathrm{fol,d}}/0.25 \approx 8.7\hat{m}_{\mathrm{fol,C}}$. (iii) Over the course of the year, m_fol relates to $\hat{m}_\mathrm{fol}$ following measured vegetation area index (VAI) variations:

$$m_\mathrm{fol} \approx \hat{m}_\mathrm{fol}\frac{\mathrm{VAI}-\mathrm{VAI}_\mathrm{max}}{\mathrm{VAI}_\mathrm{max}-\mathrm{VAI}_\mathrm{min}} \tag{9}$$

If heat storage change in the canopy air-space is given by (5), and the above assumptions are adopted, heat storage in foliage can be expressed as

$$\Delta Q_{\mathrm{v,fol}} \cong \frac{m_\mathrm{fol}}{h} c_\mathrm{veg} \sum\nolimits_{i=1}^{n} \left(\frac{\Delta (T_\mathrm{a})_i}{\Delta t}\Delta z_i\right), \tag{10}$$

where it is assumed that the foliage mass is distributed uniformly over the canopy height (h), and the specific heat of foliage is the same as that used in (7). Thus, taking ρ = 1.246 kg m^{-3}, and c_p = 1.01 × 10^3 J kg^{-1} K^{-1}, the ratio of $\Delta Q_{\mathrm{v,\ leaf}}$ to ΔQ_a is obtained as:

$$\begin{aligned}\frac{\Delta Q_{\mathrm{v,leaf}}}{\Delta Q_\mathrm{a}} &\cong \frac{m_\mathrm{fol} c_\mathrm{veg}}{h\rho c_\mathrm{p}} \cong \frac{\hat{m}_\mathrm{fol} c_\mathrm{veg}}{h\rho c_\mathrm{p}}\frac{\mathrm{VAI}-\mathrm{VAI}_\mathrm{min}}{\mathrm{VAI}_\mathrm{max}-\mathrm{VAI}_\mathrm{min}}\\ &\cong 0.15\frac{\mathrm{VAI}-\mathrm{VAI}_\mathrm{min}}{\mathrm{VAI}_\mathrm{max}-\mathrm{VAI}_\mathrm{min}}\end{aligned} \tag{11}$$

In conclusion, $\Delta Q_{\mathrm{v,\ leaf}}$ is seen to attain a maximum of 15% of ΔQ_a, in July and August (when VAI is highest), but will be considerably less during the rest of the vegetative season and zero in the leaf off period. Seeing that ΔQ_a attains about 33% of hourly ΔQ_S in July and August, this result confirms the finding of Blanken et al. (1997) that $\Delta Q_{\mathrm{v,\ leaf}}$ contributes less than 5% to the overall heat storage change flux also at the MMSF site. Thus, $\Delta Q_{\mathrm{v,\ leaf}}$ will be neglected from further analysis in this study.

Once each of the above components of the storage heat flux was derived, ΔQ_S was calculated for each hour using (2).

3. Results

3.1. Storage heat flux

Monthly ensemble averages of hourly values of each of the components of the storage heat flux are presented in Fig. 1. Seasonally, the diurnal pattern appears to be influenced by both solar radiation and

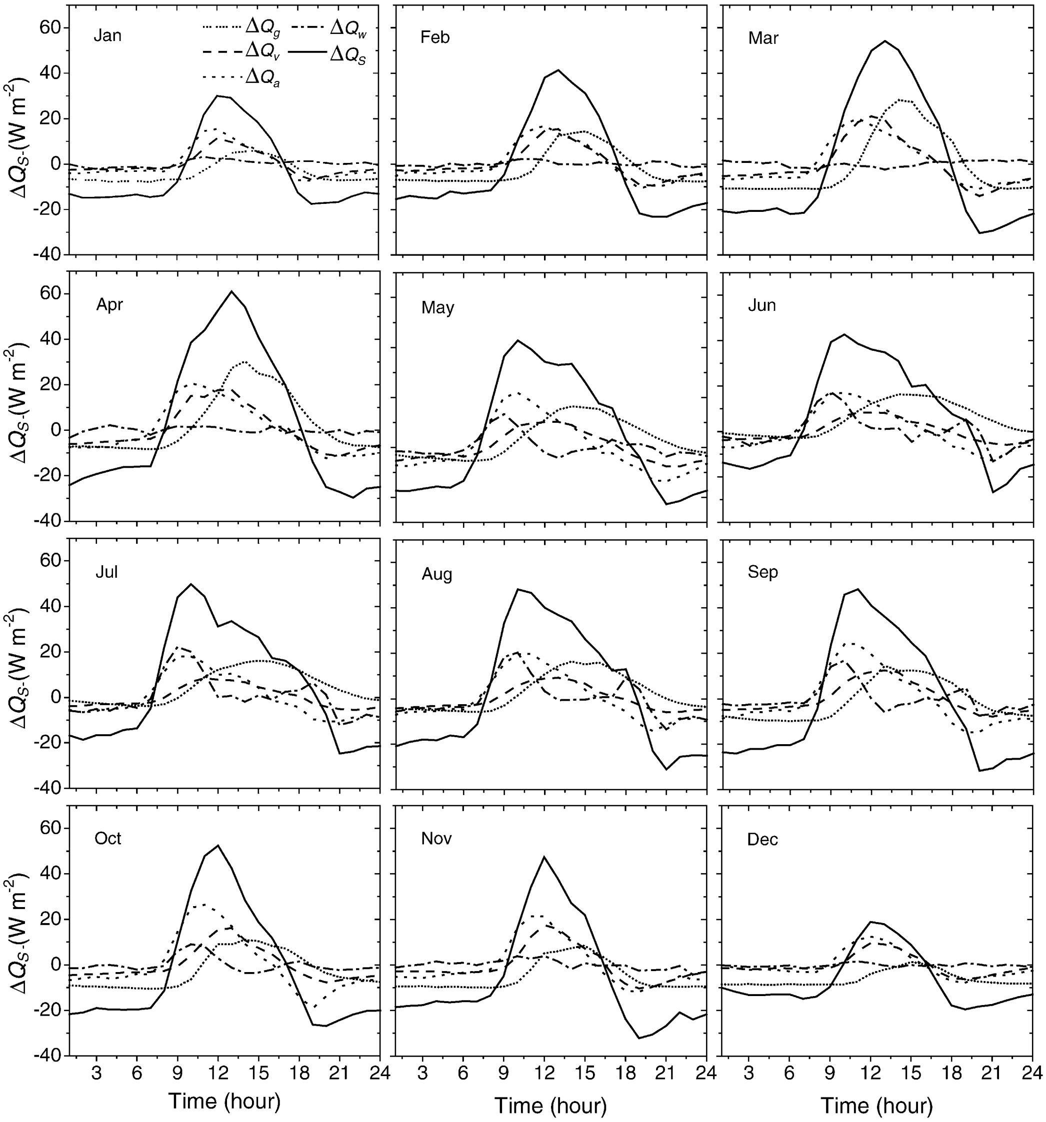

Fig. 1. Monthly ensemble averages of hourly components of the storage heat flux at MMSF from 1998 to 2001.

ecosystem processes. Storage heat flux peaks in mid-spring, just prior to leaf emergence, when insolation is the highest during the leaf-off period. Shading from leaves thereafter, reduces the diurnal magnitude of ΔQ_g, ΔQ_a, and ΔQ_v, although ΔQ_w becomes increasingly important throughout the growing season, both in hourly magnitudes and in shifting the diurnal peak to earlier in the day. Overall, ΔQ_S appears to be determined almost in equal parts by ΔQ_g, ΔQ_a, and ΔQ_v. However, when removing the diurnal signal by calculating monthly averages of daily total heat storage flux for each component, we find the seasonal shift in ΔQ_S to be most strongly dominated by ΔQ_g (Fig. 2) due to the higher thermal mass per unit area of the soil than the biomass or canopy air. For the three complete years in the dataset (1999–2001), missing hours were first gap-filled using 10-day running ensemble averages of hourly data and the annual net ΔQ_S was computed. In all 3 years, ΔQ_S showed a small annual deficit with an average of

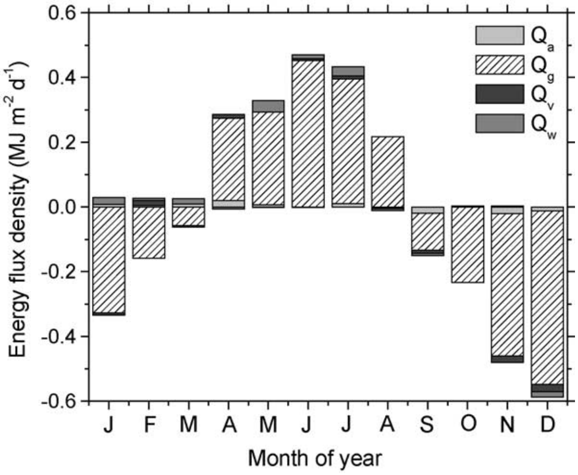

Fig. 2. Monthly averages of daily total energy contributions for components of the storage heat flux at MMSF from 1998 to 2001.

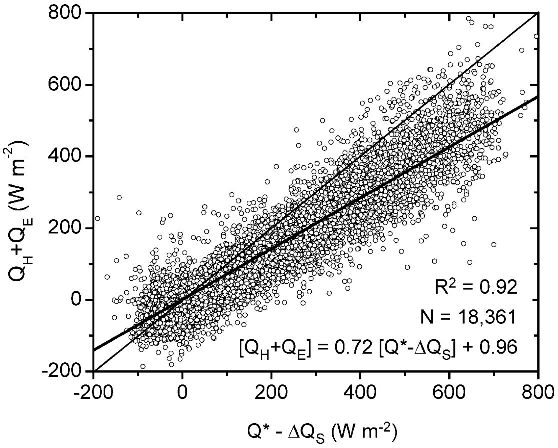

Fig. 3. Relation between turbulent heat fluxes and available energy at MMSF using hourly averages from 1998 to 2001 from 15 min eddy covariance values for those hours when all energy balance components were available.

16.18 MJ m^{-2} a^{-1}, equating to a daily average loss of 0.044 MJ m^{-2} day^{-1}. This value includes actual year to year changes in heat storage as well as unmeasured components and errors associated with sampling issues and methodological decisions. This annual residual heat storage is the same order of magnitude of MAD found using a range of c_{veg} values and is smaller by an order of magnitude to MAD from different sampling methods for vegetation and air temperature. It is also of a similar magnitude to unmeasured components of ΔQ_S ($\Delta Q_{v,\,leaf}$ and Q_p) estimated here and reported by Blanken et al. (1997).

3.2. Energy balance closure

The closure of the surface energy balance is a theoretical requirement of the first law of thermodynamics. Thus the fact that observations of the energy balance seldom close, particularly over forested sites (e.g. Wilson et al., 2002), points to insufficiencies in the experimental set-up, the sensors, or to incomplete coverage of relevant energy flux components. We examine the quality of energy flux observations at the MMSF site by determining the degree to which available energy ($Q^* - \Delta Q_S$) is balanced by turbulent fluxes ($Q_H + Q_E$), using linear regression analysis, following Wilson and Baldocchi (2000). The result of this analysis for all available hourly values from 1998 to 2001 (N = 18,361) is shown in Fig. 3, and indicates the magnitude of the turbulent fluxes to be less than that of available energy by 28%, with significant consistency. This degree of mismatch lies within the range (1–45%) of the 50 site-years of data reported in Wilson et al. (2002).

Wilson et al. (2002) discuss possible causes for the lack of energy balance, including: (i) mismatch in source areas for the terms in (1), (ii) systematic bias in instrumentation, (iii) neglected energy sinks, particularly components of ΔQ_S, (iv) the loss of low and/or high frequency contributions to turbulent fluxes, and (v) neglected advection of heat. In order to examine these possibilities, closure was investigated under varying conditions designed to evaluate the causes for lack of closure at MMSF. This was conducted using all complete sets of hourly data from 1998 to 2001. The degree of closure was determined both by linear regression of [$Q^* - \Delta Q_S$] to [$Q_H + Q_E$] for subsets of the time series and by calculating the energy balance fraction (Ω) for a given time period:

$$\Omega = \frac{\sum(Q_H + Q_E)}{\sum(Q^* - \Delta Q_S)} \tag{12}$$

where perfect closure is found when $\Omega = 1$.

The impact of mismatch in source area of different energy components is very difficult to determine given the temporal variability in turbulent flux footprint relative to the static footprint of the net radiometer and storage flux observations (Schmid et al., 1991). The diameter of the net radiometer 90%-source area on the tower is at most about 150 m (on the forest floor), centered directly beneath the instrument, while the turbulent flux footprint ranges from the same order of magnitude to two orders greater with orientation in variable wind directions. The four ground heat flux sites and six trees with thermocouple measurements are spread across up to 150 m in the prevailing upwind direction of the tower and include multiple tree species sited on different slope aspects (Section 2.3). As a rough sensitivity test of energy balance closure to footprint variations and advection, energy balance closure was tested for all wind directions (10 degree bins) and showed no statistically significant difference associated with any direction, including when the instruments were influenced by flow disturbance from the tower. This result is not surprising, as footprint mismatches or advection are unlikely to cause much long-term consistency in Ω, except in conditions where a strong directional surface inhomogeneity is associated with a dominant wind direction. Moreover, Schmid et al. (2002) indicate that the turbulent flux footprint for the 46 m measurement level at the MMSF site is statistically representative of the larger forested area to within less than 3%. Thus, turbulent flux footprint or advection issues are unlikely to contribute significantly to the energy balance mismatch found at this site.

Seasonal control on closure was evaluated both by examining Ω as a function of day of year, and through linear regression of monthly subsets, which produced no statistically significant change in closure as a function of season, also found by Wilson et al. (2002) for 50 site years. Diurnal change in closure was tested by plotting Ω against a normalized time unit where 0 is sunrise and 1 is sunset (Fig. 4) for both the complete dataset (open squares) and after rejecting values when u^* was less than 0.3 m s^{-1} (closed circles). Overall, the best closure occurs during daylight hours, peaking shortly before sunset. Nocturnal closure is relatively poor, although is significantly higher when low u^* values are rejected. In addition, closure appears to improve considerably throughout daylight hours and to a lesser extent through the night. As $[Q^* - \Delta Q_S]$ approaches zero around sunrise and sunset, Ω becomes less meaningful. Similar features were found by Wilson et al. (2002), although they used absolute time applied to a range of site latitudes and times of year.

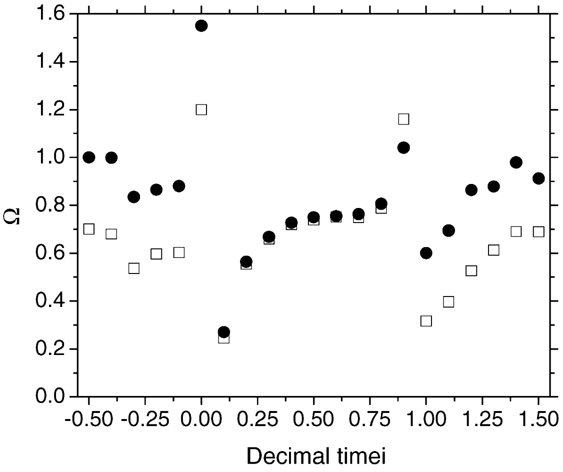

Fig. 4. Energy balance closure as a function of time of day normalized to sunrise (0) and sunset (1) at MMSF from 1998 to 2001, where open squares represent Ω calculated from the complete dataset and closed circles represent data where $u^* > 0.3$ m s^{-1}.

A realistic physical explanation for the improvement of closure during daylight hours and the general difference between day and night time regimes could be linked to the lower friction velocity (u^*) values experienced at night. Low u^* conditions are often indicative of poorly developed turbulence. Thus, they point to conditions where eddy covariance fluxes are prone to underestimate the true convective exchange, contributing to lack of closure. These conclusions are reinforced by the relation between Ω and u^* (Fig. 5a) and turbulent fluxes (Fig. 5b), illustrating that rapid decrease of Ω occurs when u^* is less than 0.3 m s^{-1} and the sums of the turbulent fluxes are close to zero. The relation described by Fig. 5a was also reported by Blanken et al. (1997) and Wilson and Baldocchi (2000). Indeed, when the mean diurnal pattern of Ω is recalculated after removing data where u^* is less than 0.3 m s^{-1}, a substantial improvement of closure (by ∼15%) is found during nocturnal hours, with very little change during the day (Fig. 4). Table 2 provides slope ratios, offsets and coefficients of determination values for linear regression as presented in Fig. 3, for each year using: (a) all data, (b) data where both available energy and the sum of the turbulent fluxes were greater than 50 W m^{-2}, and (c) where u^* is greater than 0.3 m s^{-1}. These statistics show very little

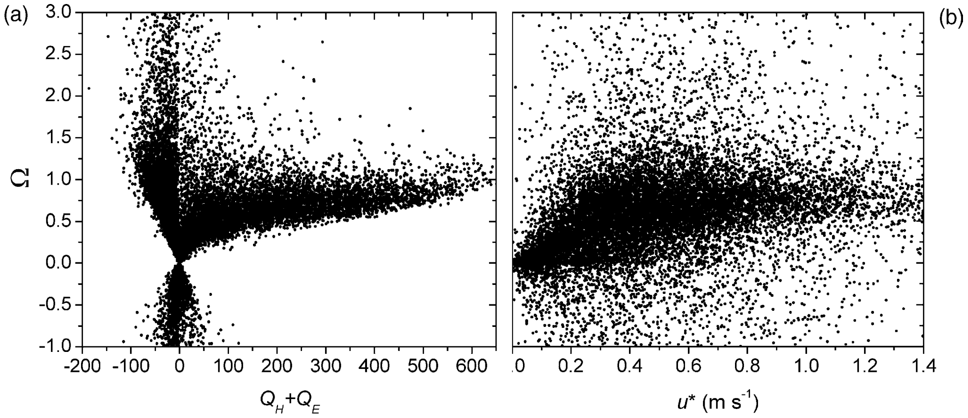

Fig. 5. Hourly energy balance closure as a function of (a) turbulent fluxes and (b) friction velocity at MMSF from 1998 to 2001. Axes have been restricted to show central tendency of data, $N = 16{,}586$.

inter-annual variability in closure performance, but modest improvement with rejection of low u^* and low energy flux values.

A commonly cited reason for the underestimation of turbulent fluxes is loss of signal from either end of the turbulent frequency spectrum (e.g. Massman, 2000; Massman and Lee, 2002). In particular, Su et al. (2004) showed a significant long-tube damping effect of water vapor on the calculation of latent heat fluxes. This analysis was conducted for the MMSF site during daylight hours of the growing season, and provided three methods for estimating flux loss (Su et al., 2004). By selecting this time period, we remove the low energy and low u^* values that were found to weaken closure performance. First, hourly fluxes derived from 15 and 60 min sampling periods were compared to assess the role of low frequency loss. Second, the role of long-tube damping of water vapor was examined by employing the corrections of Su et al. (2004), their Method-3, which determines the flux loss by the ratio of measured co-spectra to the derived transfer function integrated over the full range of frequency. Table 3 shows a small (2–3%) but inter-annually consistent improvement of closure as a function of increasing the sampling period. The co-spectral correction for Q_E led to essentially no change in the slope in 1998, but much

Table 2
Linear regression slope, offset and coefficient of determination for $Q^* - \Delta Q_S$ (independent variable) and $Q_H + Q_E$ (dependent variable) for 4 years under three conditions

Time period	1998	1999	2000	2001	All years
(a) All available data					
Slope	0.72	0.69	0.73	0.71	0.72
Offset	5.09	3.82	2.19	2.23	3.21
R^2	0.92	0.90	0.92	0.91	0.91
(b) Low energy fluxes removed					
Slope	0.74	0.73	0.77	0.76	0.75
Offset	−2.05	−7.07	−10.2	−13.9	−9.20
R^2	0.84	0.85	0.86	0.86	0.85
(c) Low u^* removed ($u^* < 0.3\ \mathrm{m\,s^{-1}}$)					
Slope	0.76	0.72	0.76	0.72	0.74
Offset	−3.02	−4.74	−5.27	−0.82	−3.17
R^2	0.92	0.90	0.92	0.91	0.91

Table 3
Linear regression slope, offset and coefficient of determination for $Q^* - \Delta Q_S$ (independent variable) and $Q_H + Q_E$ (dependent variable) for the growing season, daytime period for 4 years under three conditions

Time period	1998	1999	2000	2001	All years
(a) 15 min averaging period					
Slope	0.76	0.76	0.82	0.84	0.80
Offset	−11.2	−5.04	−19.3	−49.6	−26.0
R^2	0.77	0.72	0.76	0.79	0.76
(b) 60 min averaging period					
Slope	0.78	0.78	0.85	0.87	0.81
Offset	−12.9	−0.73	−34.6	−56.4	−31.4
R^2	0.68	0.70	0.78	0.76	0.73
(c) Co-spectral corrected					
Slope	0.78	0.86	1.03	1.09	0.97
Offset	−5.20	−3.12	−40.6	−73.9	−39.1
R^2	0.69	0.73	0.81	0.78	0.74

greater slope increases in subsequent years, from 0.08 in 1999 to 0.22 in 2001. In fact, while the co-spectral corrections appear to improve closure significantly when considering all years (slope = 0.97), the breakdown of years illustrates that improvement in closure ranges from a negligible increase in slope in 1998 to an increase by 12% in 2001. The change over time is thought to be attributable to the increase in damping due to dirtying and/or decay of tubes over time, as it occurs for water vapor, but not CO_2 (Su et al., 2004). Although considerable improvement to energy balance closure is found overall following co-spectral corrections to Q_E, it is clear that other components (measured and unmeasured) must contribute to the lack of closure. This is supported by the fact that pre-correction closure statistics did not decrease between 1998 and 2001, while the magnitude of correction increased. Furthermore, analysis of the relation between Ω and Bowen ratio ($\beta = Q_H/Q_E$) revealed no general difference in closure performance between periods when Q_H or Q_E dominated the turbulent flux throughout the observational period.

The contribution of vertical and horizontal advection in flux estimates (hypothesis v) has recently received attention, particularly for vertical advection (Lee, 1998; Baldocchi et al., 2001; Paw et al., 2000), for sites that contain topographical variability (e.g. Mahrt and Larsen, 1990; Lee, 1998; Sun et al., 1998) and sites showing strong horizontal heterogeneity in surface fluxes (Vidale et al., 1997; Leclerc et al., 2003). Although the magnitude of ΔQ_A is not known for MMSF, surface cover heterogeneity is not thought to play an important role, with no dependence of closure found on wind direction, as discussed above. However, topographical variability is considerable and Froelich and Schmid (2002) identified local thermotopographic flows in the vicinity of the MMSF tower as potential mechanisms for advection.

3.2.1. Forcing energy balance closure

To estimate total energy or mass fluxes, it is convenient to force closure by redistributing the hourly residual ($Q^* - Q_H - Q_E - \Delta Q_S$) among the measured fluxes according to some objective rule. Based on evidence presented in this section, we attribute the general pattern of imbalance to a systematic underestimation of the turbulent fluxes. Since the Bowen ratio (β) indicates the relative magnitude of each of the turbulent heat fluxes at a given time, and we find no evidence that closure changes as a function of β, it is used in this study to partition the remaining energy required to close the energy balance so that:

$$Q_{E(f)} = (Q^* - \Delta Q_S)(1 + \beta) \quad \text{and} \quad Q_{H(f)} = \beta Q_{E(f)} \tag{13}$$

where (f) refers to fluxes altered to close the energy balance. However, when fluxes are very small (within ±5 W m^{-2}) or when β approaches −1, (13) yields large diverging corrections of opposing sign. When these conditions were present (43% of all hours, mostly after sunset and before sunrise), the energy balance was closed by partitioning the residual energy equally between Q_H and Q_E. Fig. 6 shows a histogram of total residual energy for each hour that was redistributed to turbulent fluxes showing a positively skewed distribution with a mode of zero, and 66% of values falling within the range −50 to 50 W m^{-2}.

3.3. Inter-annual comparison of energy balance components

Ten-day running mean time series for each year of energy balance components (without forcing closure) and β as well as periodic observations of VAI measurements using a pair of LiCor 2000 plant canopy analyzers (LiCor, Lincoln, Nebraska) are presented in Fig. 7. Remarkable interannual consistency is found in

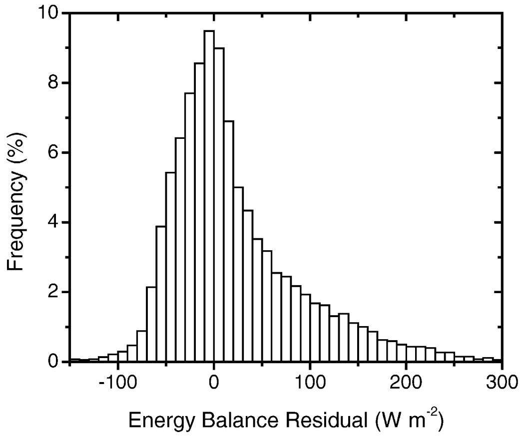

Fig. 6. Frequency histogram of residual energy in hourly energy balance observations at MMSF from 1998 to 2001.

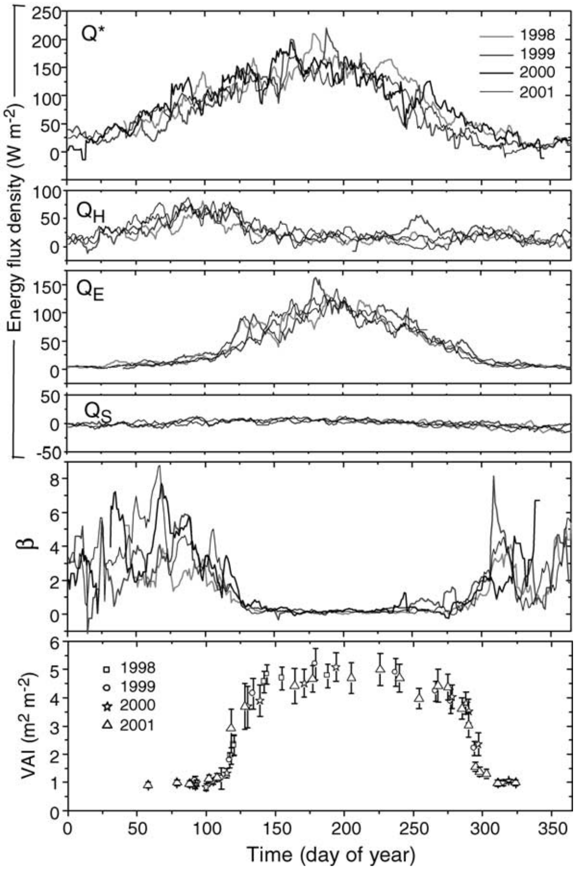

Fig. 7. Ten-day running averages for hourly components of the surface energy balance (unforced) and Bowen ratio as well as periodic average and standard deviation of 30-point VAI observations at MMSF, from 1998 to 2001.

energy balance components over the 4 years of observations. Seasonal controls are clearly evident, both in terms of the net radiant energy input, which peaks near the summer solstice and in terms of phenological changes in the forest. Q_H peaks in spring, at the time of year when net radiation is highest during leaf-off, with a smaller secondary peak following leaf senescence. As leaves emerge around day 110, a simultaneous drop in Q_H and rapid jump in Q_E occurs as transpiration is initiated. Q_E then peaks in concert with Q^*. Plotted on the same scale, ΔQ_S can be seen to make a very small contribution to the overall energy balance. β is seen to respond closely to phenological controls appearing inversely related to VAI, with dual peaks either side of the growing season. The filtering (10-day running mean) in Fig. 7 removes the diurnal signal, but captures significant variability on the time scale of days to weeks, reflecting the influence of broader synoptic patterns.

The most notable interannual difference appears during the late summer (days 240–270) of 1999, following a period of anomalously low rainfall and low soil moisture (not shown), when Q_H was relatively dominant and β was higher than other years. However, this response in energy partitioning was relatively small suggesting the effectiveness of deep roots in sourcing water stored in the soil column. Four years of VAI data show little change in the timing of leaf emergence (average initiation date = day 99) and senescence (average completion date = day 301). Spatial variability in the magnitude of VAI is indicated by the vertical bars in the Figure, which represent the standard deviation of observations sampled at 30 fixed locations. Direct observations of phenology provide more precise information about leaf emergence (at least 50% of leaves expanded on all trees observed): day 100 in 2001, but later on day 117 or 118 in 1998, 1999, and 2000. However, the inter-annual effects on Q_H and β are less apparent.

Power spectra for energy balance components for the 4-year data set show the expected major peaks at the semi-diurnal, diurnal, semi-annual and annual timescales (not shown). The semi-annual peak is strongest in Q_H (almost equal to the annual peak) and absent in ΔQ_S. This is thought to result from the secondary seasonal influence of biophysical controls evident in Fig. 7, which impacts annual patterns of Q_H the greatest. Smaller and less consistent peaks are evident on the frequency of days to weeks, suggestive of broader synoptic controls but these will be underestimated given the gap-filling strategy used.

For the three full operational years (1999–2001), annual cumulative energy was calculated. First, gaps in the observational record for each of the energy balance components were filled using 10-day running ensemble average data and the energy balance was forced to close for each hour. Fig. 8 shows that interannual variability in cumulative flux between the 3 years is relatively small. Among the three largest components, interannual variability remains relatively

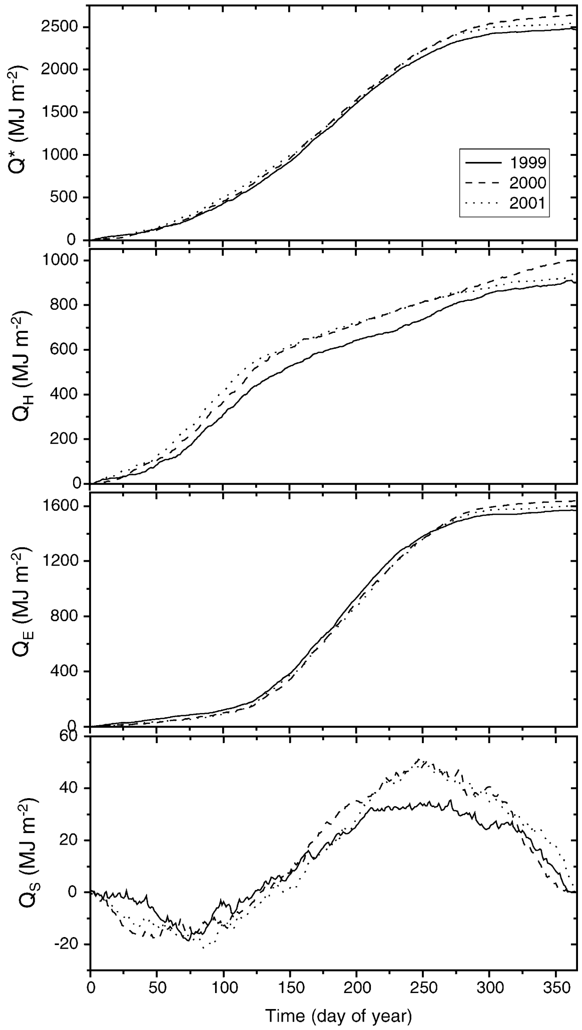

Fig. 8. Daily total cumulative energy fluxes for components of the surface energy balance after forcing closure at MMSF, from 1999 to 2001.

small, in terms of timing, rates of change, magnitudes throughout the year and the annual cumulative totals (Table 4). The annual β averaged 0.591 with a range of only 0.044. The largest absolute variability in flux totals was found in Q^*, although as a percentage of the 3-year average, Q_H showed the largest difference, with a maximum of 5.96%.

Comparison of MMSF energy balance totals with values reported from a mixed deciduous forest in Oak Ridge, Tennessee (Wilson and Baldocchi, 2000) revealed that, despite somewhat greater magnitudes of all fluxes (Oak Ridge is 3.5° closer to the equator), the relative magnitudes of all fluxes and of β are similar. In addition, over the reported observation period (1995–1997) Oak Ridge experienced similarly small but slightly greater interannual variability than MMSF (up to 8% difference for individual years from the 3-year average).

4. Conclusions

Based on 4 years (1998–2001) of hourly forest-atmosphere energy exchange observations, this study has presented components of the surface energy balance for the MMSF site. An emphasis was placed on estimating the sub-components of the storage heat flux term (ΔQ_S). The diurnal pattern of ΔQ_S had significant contributions by ΔQ_g, ΔQ_v and ΔQ_a, but ΔQ_g was the only component to show significant storage changes on a seasonal scale. The role of phenology on ΔQ_S was also evident, with a reduction of many of the components during the growing season except ΔQ_w, which increased. The annual average ΔQ_S budget was 16.18 MJ m^{-2} a^{-1} in deficit. Distributed over the year, this deficit is smaller than mean absolute differences generated

Table 4
Annual Bowen ratio and totals of energy observed at MMSF and the annual departure from the average of the 3 years (ΔAv) expressed as a percentage

Year	β	Q^*		Q_H		Q_E	
		Total (MJ m^{-2})	ΔAv (%)	Total (MJ m^{-2})	ΔAv (%)	Total (MJ m^{-2})	ΔAv (%)
1999	0.57	2469	−3.35	899	−5.40	1570	−2.16
2000	0.614	2647	+3.61	1007	+5.96	1640	+2.20
2001	0.589	2548	−0.26	945	−0.56	1604	−0.04

from sensitivity tests of various methodological constraints. In particular, under-sampling of biomass temperature and relative humidity have a significant impact on ΔQ_v and ΔQ_w respectively.

Energy balance closure at MMSF for 4 years, determined by linear regression, found turbulent fluxes ($Q_H + Q_E$) to underestimate available energy ($Q^* - \Delta Q_S$) by 28 % with a small offset (0.96 W m^{-2}) and with considerable consistency ($R^2 = 0.92$, $N = 18{,}361$). No statistically significant difference in closure was found among years or among months. Conversely, closure (Ω) is linearly reduced with decreasing mixing when $u^* < 0.3$ m s^{-1}. Quality of closure was also found to improve as a function of increasing magnitude of turbulent heat fluxes. Comparison of 15 and 60 min block averaging times for turbulent flux calculations indicated a small improvement in closure (2.5%) using the longer period. Co-spectral corrections to Q_E for long-tube damping effects generated a much larger improvement in closure (16%). Variability of closure was not found to relate to wind direction, suggesting that footprint mismatches or advection do not significantly impact energy balance closure at this site.

Time-series analysis revealed strong Sun–Earth geometric controls on energy balance in terms of available energy, governed most strongly at the diurnal and seasonal timescales, and strong biophysical controls on energy partitioning, particularly β, which showed an abrupt shift from Q_H to Q_E dominance of available energy at the onset of the growing season. The period just prior to leaf emergence also marks the seasonal peak of Q_H, the period in which peak available energy and β coincided. During senescence, β shifted back to larger integers, showing the return of Q_H dominance. Temporal variability at the timescale of days to weeks was also significant, reflecting temporal patterns in synoptic controls. Interannual variability of energy balance components was remarkably small, both in magnitude and relative partitioning as well as phenological patterns. The mean annual Bowen ratio was 0.59, ranged from 0.57 to 0.61 and inter-annual variability of all energy balance components over the period was less than 6% departure from mean. Further work is required to evaluate the causes for inter-annual similarity in this case, both in temporal context of the longer climate record and the spatial context of comparison with other ecosystems.

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