

Seasonal and interannual variability of energy fluxes over a broadleaved temperate deciduous forest in North America

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Abstract

The components of the surface energy balance were measured for 3 years over a broadleaved deciduous forest using the eddy covariance technique. Within years, the magnitude and distribution of fluxes was controlled by seasonal changes in solar radiation, drought, as well as leaf emergence and senescence. Evapotranspiration increased by a factor greater than five (from about 0.5 to 3 mm day⁻¹) after leaves emerged in spring. Large decreases in sensible heat flux were observed over the same period (6 to 2 MJ day⁻¹) despite increases in solar radiation. The most influential effect on annual fluxes was the occurrence and extent of drought, with lesser control exerted by differences in the timing of leaf expansion and leaf senescence. Average annual evapotranspiration over the period was 567 mm and ranged from 537 to 611 mm. The year with the lowest precipitation, soil moisture content and surface conductance also had the lowest evapotranspiration. Although evapotranspiration was quite sensitive to surface conductance and surface conductance was reduced substantially by drought, the correlation of low surface conductance and high humidity deficit reduced the effects of drought on evapotranspiration. Differences in net radiation among years were only a minor source of variability in evapotranspiration. In addition to surface conductance, other bulk parameters are calculated to describe the general exchange characteristics of this forest. ©2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

Deciduous forests cover much of the eastern United States. Meteorologists, hydrologists and forest ecologists require knowledge of how these forests and the atmosphere exchange energy and water. Energy parti-

tioning, particularly between latent and sensible heat, determines the water vapor and heat content of the atmosphere, ultimately driving many regional and global scale climatological processes (Dirmeyer, 1994; Seth and Giorgi, 1996). This partitioning determines the growth rate and properties of the planetary boundary layer, influencing convection, long range transport of heat, humidity and pollutants. Evapotranspiration also affects streamflow, nutrient loss, soil moisture content and forest productivity.

An understanding of the variability of energy and water vapor fluxes over a site requires long-term measurements encompassing at least several different

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climatic and phenological regimes creating this variability. Deciduous forests are characterized by phenological stages including leaf emergence and senescence (see Hutchison et al., 1986), which can vary in timing and duration between years and alter the exchange properties between the forest and atmosphere. The effect of these events within a year and the timing of these events between years on long-term fluxes require high-resolution and continuous extensive data sets. Interannual climate variability, most notably the occurrence, duration and extent of drought, can reduce water and carbon dioxide exchange (Baldocchi, 1997). The biological and climatic controls of these long-term drought episodes on forest-atmosphere exchange processes also need to be established.

The eddy covariance technique (Baldocchi et al., 1988) provides a tool which allows for direct flux measurements at the time and spatial scales required to examine biological and climatological mechanisms creating variability. A number of short campaigns over a deciduous forest in the southeastern USA have reported canopy-scale water and energy exchange with the atmosphere using this technique. These studies have shown the relative magnitude and variation that can occur over several days (Verma et al., 1986; Baldocchi and Vogel, 1996), a single season (Baldocchi, 1997) and a single year (Greco and Baldocchi, 1996). Other long-term energy balance studies based on eddy covariance at deciduous forests were single-year studies during BOREAS (Blanken et al., 1997) and at Harvard forest, Massachusetts (Moore et al., 1996). Multi-year studies of interannual variability of the energy balance are scarce.

Several benefits exist for using long-term eddy covariance data sets to evaluate ecosystem exchange. One is to validate process-based ecosystem exchange models operating at hourly, daily, seasonal and annual time scales. A second benefit, and the one developed here, is to identify and quantify the general characteristics of variability and canopy exchange processes that can be compared with other climates and biomes. Seasonal and annual sums of fluxes and the calculation of bulk canopy parameters can be used to characterize canopy exchange and estimate the controls on that exchange. Some of these proposed bulk parameters include surface conductance, the Priestly–Taylor coefficient (Priestly and Taylor, 1972), the Bowen

ratio, and the decoupling coefficient (Jarvis and McNaughton, 1986). Long-term studies establish the behavior of these parameters, and whether they can be used to characterize vegetation types over seasonal and interannual time scales. Evaluation of these parameters may also be useful in refining boundary conditions of weather forecasting, hydrological and climate models, especially if they are shown to scale with more fundamental ecosystem parameters such as leaf area, leaf nitrogen and precipitation (Baldocchi and Meyers, 1998).

In this study, we report on 3 years of eddy covariance measurements of energy and water fluxes over a temperate broadleaved deciduous forest in the southeastern United States. A summary of the components in the net energy balance will be discussed within and between the 3 years. Although diurnal and day to day variability is present in the data, the focus will be on seasonal and interannual variability. We primarily emphasize the distribution of sensible and latent heat fluxes and the possible biological and climatological controls on this distribution. Implications of variability in leaf phenological patterns and climate variability, especially drought, will be discussed. Bulk parameters, such as surface conductance, the decoupling coefficient, the Priestly–Taylor coefficient and Bowen ratio are shown to characterize canopy exchange processes and how the controls of these processes vary at seasonal and interannual time scales. The three-year data set is believed capable of characterizing at least some of the typical interannual variability at this site because of the 1995 drought compared to the wetter years of 1996 and 1997 and differences in leaf emergence amongst the years.

2. Materials and methods

2.1. General forest characteristics

Micrometeorological and flux measurements were made above a temperate deciduous forest in Oak Ridge TN (35°57'30" N, 84°17'15" W, 365 m asl) continuously from 1995 through 1997. The site is located in the southern section of the temperate deciduous forest biome in the eastern United States. The canopy

height was approximately 26 m above the surface. The forest contains a mixed deciduous stand dominated in the overstory by oak, maple and hickory. The stand is over 50 years old, having regenerated from agricultural land. The upwind fetch of forest extends several kilometers in all directions. The soil is well drained and is classified as a typic Paleudult, which encompasses clayey and kaolinitic soils. A more detailed description of the canopy architecture, species composition and soil properties are provided by Peters et al. (1970), Luxmoore et al. (1981), Hutchison et al. (1986) and Johnson and van Hook (1989).

2.2. Flux system and meteorological instruments

The instruments designed to measure fluxes were placed on a scaffold tower 36.9 m above the surface and about 10 m above the canopy. Wind velocity and virtual temperature fluctuations were measured with a three-dimensional sonic anemometer (model SWS-211/3K, Applied Technology, Boulder, CO). Fluctuations in water and CO₂ were measured with an open path, infrared absorption gas analyzer (Auble and Meyers, 1992), which was calibrated monthly using gas standards prepared by NOAA's Climate Monitoring and Diagnostic Laboratory. Water vapor calibrations were referenced to a dew point hygrometer.

Simultaneous to the flux measurements, environmental and meteorological variables were measured at 1 s intervals and logged on digital data loggers (model CR-21x, Campbell Scientific, Logan Utah). Soil heat flux density was measured with three soil heat plates [model HFT-3, Radiation Energy Balance systems (REBS), Seattle, Washington] buried 0.01 m below the soil surface. Air temperature and relative humidity were measured with a temperature/humidity probe (HMP-35 A, Vaisala, Helsinki, Finland). Photosynthetically active radiation (PAR) was measured above and below the canopy with a quantum sensor (model LI-190S, Licor Inc., Lincoln, NE). The sensor below the canopy was placed on a moving tram to average PAR over a horizontal transect of 20 m. Net radiation above the canopy was measured using a net radiometer (model 7, REBS, Seattle, Washington). Canopy bole temperature was calculated using three thermocouple probes inserted 1 cm into the trunk of a

tree at breast heat. Soil was periodically collected for gravimetric measurements of soil water content.

2.3. Data processing

Vertical flux densities were evaluated by computing the mean covariance of water and sensible heat fluctuations with the fluctuating vertical velocity (Baldocchi et al., 1988). Fluctuations of velocity and scalars from the mean were determined from the difference between the instantaneous values and the mean scalar quantities. Mean scalar values were determined using a digital recursive filter with a 400 s time constant. Coordinate axes were rotated so that the mean vertical velocity was zero (McMillen, 1988). Water vapor and carbon dioxide fluxes were corrected for the effect of density fluctuations (Webb et al., 1980).

Canopy heat storage (C_s) was approximated from $M_a C_H \Delta T/t$, where M_a is the biomass per unit ground area (kg m^{-2}), C_H is the specific heat of the combined woody biomass and water ($3.34 \text{ kJ kg}^{-1} \text{ C}^{-1}$), ΔT is the change in bole temperature at 1 cm depth (C) and t is the time (s) over which ΔT is calculated. In 1996 and 1997 maximum leaf area was estimated from litter collection baskets. Total plant area index was computed continuously by applying Beers Law to solar radiation measurements above and below the canopy (Greco and Baldocchi, 1996). Leaf area was obtained by subtracting the woody canopy area index, which was obtained during leafless periods (Hutchison et al., 1986), from the total plant area index. Volumetric soil moisture was determined from the gravimetric measurements using a bulk density of 0.9 kg m^{-3} (Peters et al., 1970).

Data were screened vigorously for anomalous turbulent statistics and sensor malfunction, which introduced periods with missing data. To obtain annual sums it was necessary to fill in missing data. Hourly latent heat fluxes (LE) that were missing or of insufficient quality were assessed from the product of equilibrium evaporation for the hour [$LE_{eq} = \varepsilon/(\varepsilon + 1)(R_n - G - C_s)$] and the 2-week average Priestly–Taylor coefficient ($\alpha = E/E_{eq}$). E is measured evaporation ($\text{kg m}^{-2} \text{ s}^{-1}$), ε is sL/Cp , where s is the slope of the saturation specific humidity versus temperature (K^{-1}), L is the latent heat of vaporization (J kg^{-1}), Cp is the specific heat capacity ($\text{J kg}^{-1} \text{ K}^{-1}$),

R_n is the net radiation (Wm^{-2}), G is the soil heat transfer (Wm^{-2}) and C_s is the canopy heat storage (Wm^{-2}). In solving for α , E and E_{eq} are the total evaporation and total equilibrium evaporation during the 2-week period when E was available. Missing sensible heat fluxes were assessed from the regression equation describing the energy balance (see Section 3.4). Much of the missing data occurred at night or when precipitation or dew obscured sensor optics, which are periods when fluxes were expected to be small. We found that monthly and annual sums of latent and sensible heat flux were not highly sensitive to whether we subjected the data to acceptance criteria that was strict, as adapted in this paper, or loose.

Surface conductance (G_c) was computed each hour by inverting the Penman–Monteith equation using the Bowen ratio (Shuttleworth et al., 1984):

$$\frac{1}{G_c} = \frac{(\epsilon\beta - 1)}{G_a} + \frac{\rho D}{E} \quad (1)$$

where ρ is the density of air (kg m^{-3}), D is the specific humidity deficit of the air above the canopy (kg kg^{-1}), β is the ratio of sensible to latent heat flux (H/LE). The aerodynamic conductance (G_a) was determined from surface layer similarity (Brutsaert, 1982). There is an ‘excess’ resistance for scalars relative to momentum, which was calculated from an empirical relationship with friction velocity (Thom, 1972). Surface conductance values for an integrated period of time were evaluated by weighting the surface conductance over a given time by the net radiation:

$$G_c \frac{\int G_c R_n dt}{\int R_n dt} \quad (2)$$

This procedure was chosen over simple time averaging because very small conductances early in the morning and late in the afternoon have small effects on integrated fluxes but can arbitrarily and substantially influence average values. Instead (2) was chosen to give greater weight to values of surface conductance during midday when fluxes were greater, without completely neglecting morning or evening data.

3. Results and discussion

A major source of intra- and inter-annual variability of energy fluxes is climatology. We first examine the

general climatic characteristics over the 3 years to set the stage for how fluxes respond to these variations.

3.1. Meteorological data

Thirty-year average annual precipitation for the site is 1372 mm. Over the study period the annual sums were 1245 mm (1995), 1682 mm (1996) and 1435 mm (1997) (Table 1). Average monthly precipitation for the 3 years is shown in Fig. 1a. In 1995 precipitation was only about 10% below normal, but the total precipitation during the critical 3-month period of June, July and August 1995 was particularly low (165 mm compared to the average 343 mm for these 3 months). Over the same period in 1996 and 1997 the total precipitation was above average (413 and 558 mm).

Fig. 1b shows the mean monthly temperature at 36 m, 10 m above the canopy. Thirty-year means are not available at this height above the canopy, but at a nearby climatological site (Oak Ridge, NOAA/ATDD), the annual departures from the 30-year normal (13.9°C) were $+1.4^\circ\text{C}$ (1995), $+0.6^\circ\text{C}$ (1996) and $+1.0^\circ\text{C}$ (1997) (Table 1). During the 3-month period of June, July and August the departures from normal were $+2.4^\circ\text{C}$ (1995), $+1.6^\circ\text{C}$ (1996), and $+1.0^\circ\text{C}$ (1997). The monthly average daytime (900 EST–2000 EST) specific humidity deficit at 36 m is shown in Fig. 1c. Along with the decrease in precipitation, the humidity deficit was greater in 1995, especially in early spring and again in mid-summer. Total solar irradiance for each of the years was within 1% (Table 1).

3.2. Soil moisture

Volumetric moisture content of the top 15 cm of soil is shown for the three seasons in Fig. 2. Throughout most of the spring and summer, soil moisture was lower during 1995 than in 1996 and 1997, coinciding with the lack of precipitation during that period (Fig. 1a). The driest period of the 3 years occurred between days 190 and 250 in 1995. In 1996 and 1997 water contents were much closer to the late winter (field capacity) values through a larger portion of the spring and summer. In 1997 soil water content

Table 1

Summary of climatology, energy balance components and bulk canopy characteristics for the 3 years. Three-year averages are shown along with departure from 30-year normal in parentheses when applicable. ‘% flux data coverage’ is the percent of hours which contain measured values for all components of the energy balance. ‘Evaporation measured’ is evapotranspiration directly measured by the eddy-covariance system, while ‘Evaporation estimated’ is the sum of the measured evaporation and the estimated evaporation when data was missing. The three values of α refer to values calculated either for (i) all hours of the year, (ii) all hours during midseason or (iii) daytime hours only during midseason. Midseason refers to the period between days 140 and 275. Dry surface conductance (G_c) and Ω are shown both for the midseason period and for the year

Variable	Year			
	1995	1996	1997	3-yr Avg
Solar radiation (GJ m^{-2})	5.45	5.43	5.41	5.43
Temperature ($^{\circ}\text{C}$)	15.3	14.5	14.9	14.9 (+1.0)
Precipitation (mm)	1245	1682	1435	1454 (+82)
Maximum leaf area index	NA	5.5	6.0	5.75
% flux data coverage	75%	81%	77%	78%
Net radiation R_n (GJ m^{-2})	3.17	2.93	3.01	3.04
Latent energy flux LE (GJ m^{-2})	1.31	1.36	1.50	1.39
Sensible heat flux H (GJ m^{-2})	1.11	0.97	1.08	1.05
Canopy heat storage C_s (GJ m^{-2})	0.00	0.00	0.00	0.00
Soil heat transfer G (GJ m^{-2})	0.02	0.00	0.01	0.01
Evaporation measured (mm)	446.7	491.5	518.9	485.7
Evaporation estimated (mm)	536.8	553.7	611.1	567.2
Equilibrium evaporation E_{eq} (mm)	876.0	793.4	834.9	834.8
α 24 h (year)	0.68	0.72	0.77	0.72
α 24 h (midseason)	0.71	0.83	0.85	0.80
α daytime (midseason)	0.64	0.76	0.76	0.72
G_c midseason ($\text{mmol m}^{-2} \text{s}^{-1}$)	0.291	0.435	0.441	0.389
G_c year ($\text{mmol m}^{-2} \text{s}^{-1}$)	0.224	0.298	0.320	0.281
Bowen ratio (year)	0.78	0.68	0.71	0.72
Ω (year)	0.22	0.28	0.28	0.26
Ω dry canopy (midseason)	0.35	0.45	0.44	0.41
Spring (day CO_2 flux changes sign)	105	121	108	111
Autumn (day CO_2 flux changes sign)	305	315	314	311
Est. growing season length (days)	200	194	206	200
Max leaf nitrogen (mg g^{-1})	NA	23.0	21.0	22.0

declined after day 210, but this decline began much later and was generally not to the extent of the decline in 1995.

3.3. Leaf area and length of growing season

Calculations of leaf area based on litter basket collection indicated maximum leaf areas of 5.5 in 1996 and 6.0 in 1997. Continuous measurements of leaf area based on Beers Law were slightly less than collected in the litter baskets, but demonstrate the dynamics of leaf area over the 3 years (Fig. 3). The emergence of leaves in spring began about 2 weeks earlier in 1995 and 1997 compared to 1996 (Fig. 3). A cool cloudy period in 1997 delayed the date of full leaf expan-

sion to nearly that in 1996. The beginning (spring) and ending (autumn) days with physiologically active leaves and total length of growing season were also estimated from the approximate dates when the daily flux of carbon dioxide crossed the zero axis in spring and again in autumn (Table 1).

3.4. Energy balance

The performance of the eddy covariance measurement system is often evaluated by examining energy balance closure (Baldocchi et al., 1988). For each half-hour period, the total net radiation (R_n) should approximately balance the sum of energy distributed between latent (LE) and sensible (H) heat flux, soil

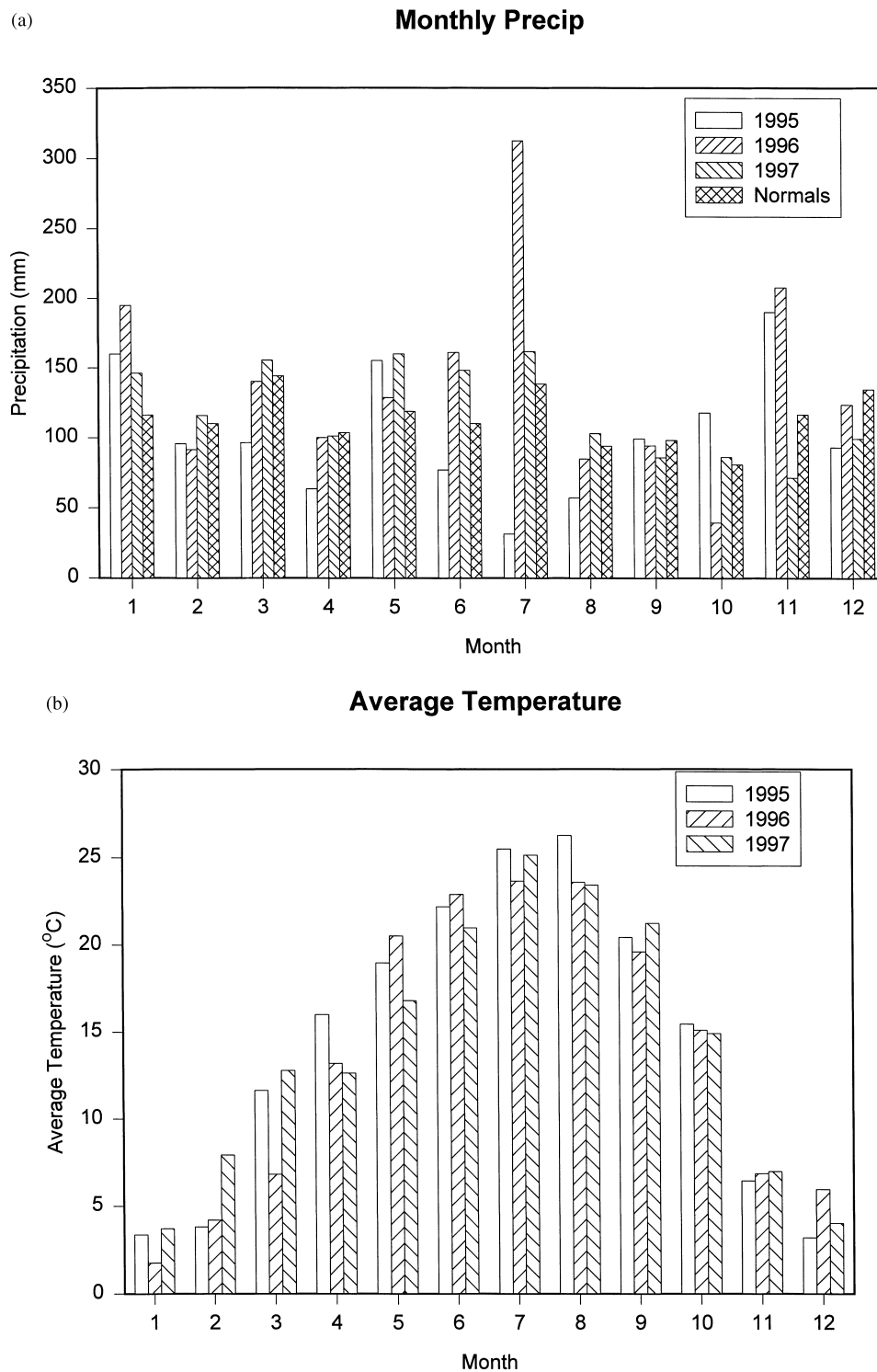


Fig. 1. Monthly averages of (a) precipitation, (b) air temperature at 36 m and (c) daytime (09:00–20:00 EST) specific humidity deficit at 36 m for each of the 3 years. Also shown in (a) are the 30-year normals.

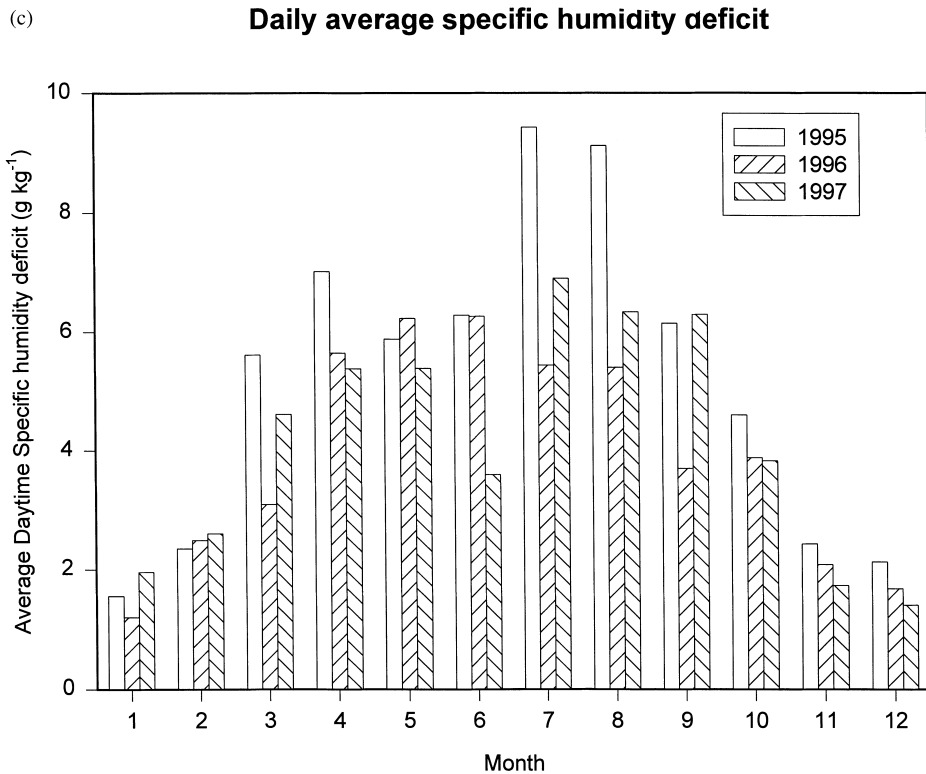


Fig. 1 (Continued).

heat transfer (G) and heat storage within the canopy (C_s) (i.e. $R_n = LE + H + G + C_s$). Regression statistics of $LE + H$ on $R_n - G - C_s$ for half-hourly data are shown in Table 2 for each of the 3 years. For all years, the slope of the regression was less than one and the intercept was slightly greater than zero. On an annual basis the sum of $LE + H$ was about 80% of the net radiation. Recent modification by REBS of the Q7 sensor calibration resulted in poorer energy balance closure compared to previous studies at this site using a Q6 (Verma et al., 1986; Baldocchi and Harley, 1995). Other long-term eddy covariance studies indicate that lack of energy balance closure is common, although this discrepancy is not well understood (Goulden et al., 1997; Blanken et al., 1997; Aubinet et al., 1999). At our site, we found no effect of wind direction on energy balance closure nor did we find evidence of flux divergence (not shown). While our site is in a region of undulating terrain, typically during the daytime wind flow is

along a broad ridge and the vertical angle of rotation is small.

3.5. Net radiation

The annual net radiation was 3.17 GJ m^{-2} (1995), 2.93 GJ m^{-2} (1996) and 3.01 GJ m^{-2} (1997) (Table 1). These annual values are greater than those reported over a 20-year period in a German deciduous forest (48°N) (Jaeger and Kessler, 1997) and for a nearby deciduous forest in West Virginia (39°N) (Tajchman et al., 1997). Although annual sums exceed those in boreal regions, mid-growing season averages of net radiation are only slightly less than for boreal sites (Baldocchi and Vogel, 1996).

Fig. 4a shows the 2-week averages of net radiation for the three seasons. Besides the obvious seasonal trend, relatively high values occurred in the spring of 1995 and low values occurred in the late spring of

Soil Moisture

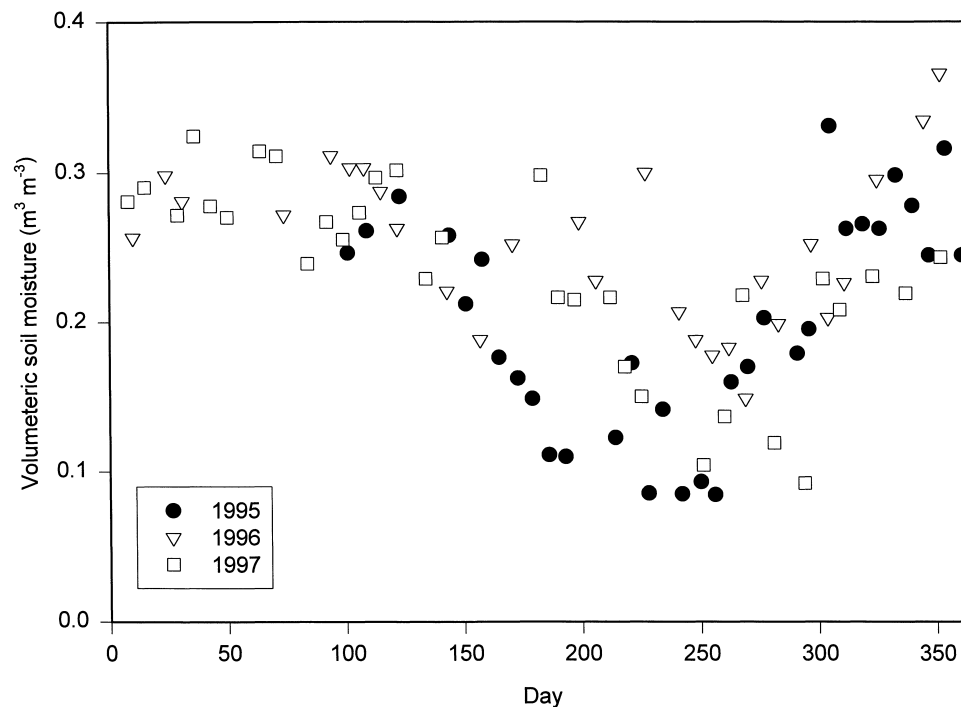


Fig. 2. Volumetric soil moisture content for the 3 years.

1997, both periods coinciding with anomalies in solar radiation. In the following sections the four energy sinks which approximately balance net radiation are discussed.

3.6. Canopy heat storage and soil heat transfer

The canopy heat storage term (C_s) was occasionally an important sink or source of energy, with a typical daily maximum value of 60 Wm^{-2} in late spring, but because the term usually changed signs between daylight and nighttime hours, the average daily magnitude rarely exceeded $2\text{--}3 \text{ Wm}^{-2}$. As expected, on annual time scales C_s was negligible and not statistically different from zero.

There was a seasonal trend in soil heat transfer (G) each of the 3 years (Fig. 4b). The soil served as an energy sink ($G > 0$) from early spring through most of the summer. The sink maximum generally occurred between mid-April and late June at an average 24 h

flux density of around 10 Wm^{-2} . Average midday rates were typically around $25\text{--}30 \text{ Wm}^{-2}$ during this period, occasionally approaching maximum hourly values of $60\text{--}70 \text{ Wm}^{-2}$, but on an hourly scale soil heat flux was usually less than canopy heat storage. Generally, from mid-morning to late afternoon during the warm season, the magnitude of the soil heat flux was less than 10% of net radiation. By late summer the soil heat flux reversed signs and reached minimum daily average values of around -10 Wm^{-2} between November and January. At night, and especially during winter, G was often a considerable portion of the negative net radiation. Although the soil heat flux was important on diurnal and seasonal time scales, the yearly-integrated soil heat flux was less than 1% of the annual net radiation (Table 1).

3.7. Sensible and latent heat fluxes

Total evapotranspiration for the 3 years was 537 mm (1995), 553 mm (1996) and 611 mm (1997). The

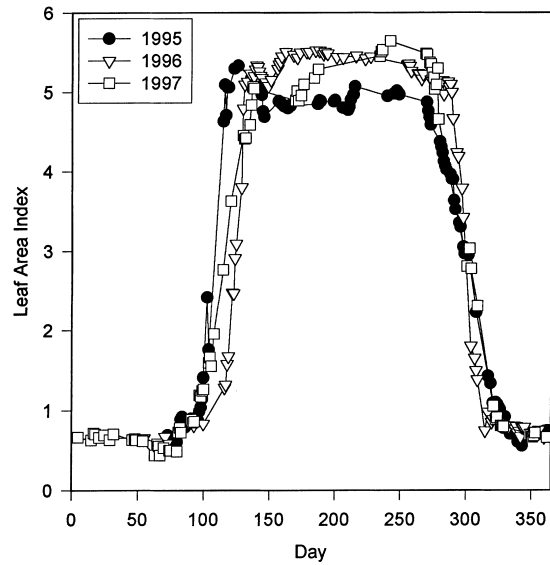


Fig. 3. Leaf area index as estimated using Beer's Law for the 3 years.

3-year average evapotranspiration for a 150-day period (day 150–day 300) during the growing season was about 33% greater than for the same period in a northern deciduous forest in Massachusetts (Moore et al., 1996). The estimated annual evapotranspiration over a deciduous aspen stand in Canada during 1994 was 403 mm (Black et al., 1996). Typical daily maximum rates ($4\text{--}5\text{ mm day}^{-1}$) were similar to those found over the aspen stand and a tropical forest (Shuttleworth et al., 1984) and about twice those over boreal coniferous forests (Kelliher et al., 1997; Baldocchi et al., 1997; Jarvis et al., 1997).

Fig. 5 shows the 2-week average daily integrated sensible and latent heat fluxes for each year. Sensible heat flux reached a maximum just before leaf emergence each year. Just after leaf emergence there was a dramatic decrease in sensible heat, while latent heat fluxes increased in an equally dramatic fashion and

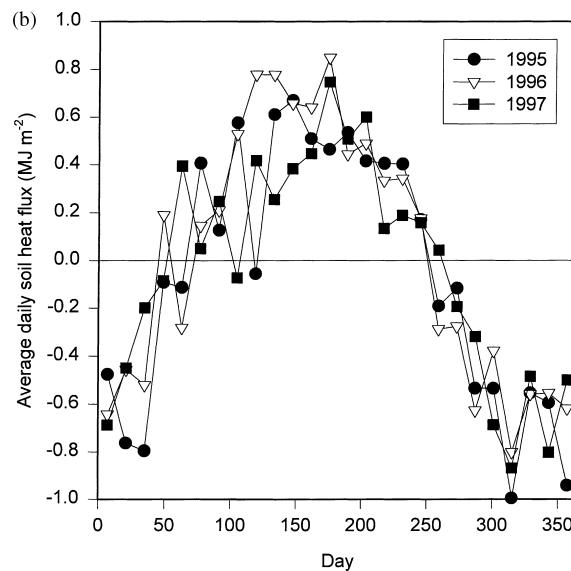
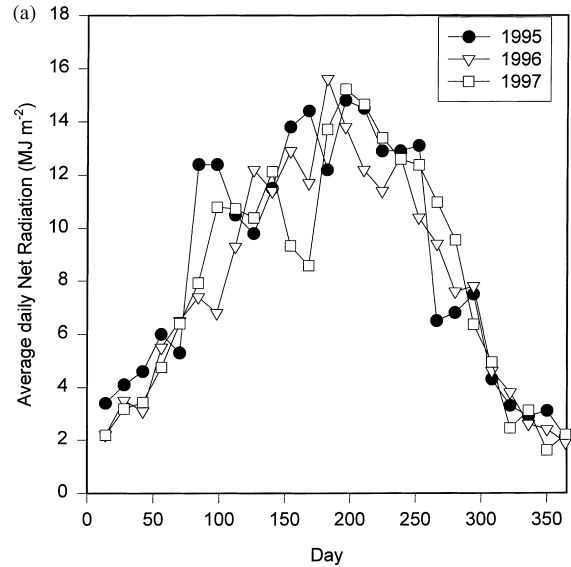


Fig. 4. Two-week averages of the daily (a) net radiation and (b) soil heat flux for the 3 years. 'Day' on the x-axis is the center point of the 2-week average.

Table 2

Slope, intercept and R^2 of the half-hourly energy balance closure ($LE + H$ versus $R_n - G - C_s$) for each of the 3 years

Year	Slope	Intercept (Wm^{-2})	R^2
1995	0.71	1.39	0.94
1996	0.76	2.21	0.95
1997	0.77	7.15	0.94

peaked from early June (1995) to early July (1997). Within about a 6–8 week period each year the sensible heat flux went from a distinct yearly maximum to a near minimum. The daily-integrated sensible heat flux was fairly constant after leaf expansion.

The distribution of energy fluxes from primarily sensible to latent heat was directly related to the

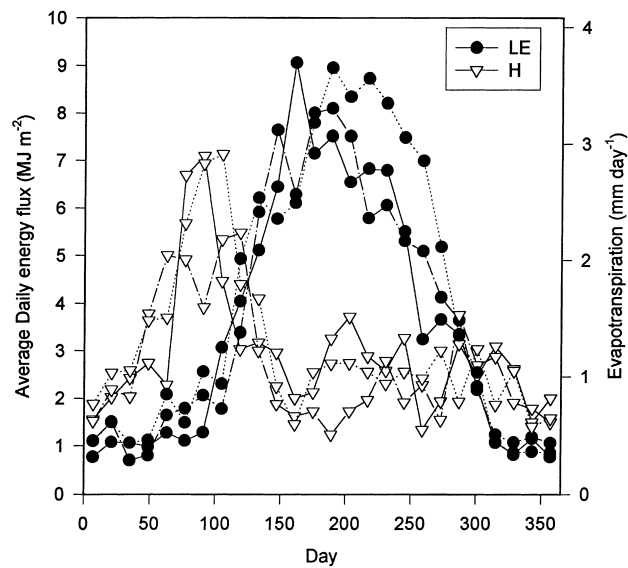


Fig. 5. Two-week daily averages of daily latent (LE) and sensible (H) heat flux for 1995 (solid line), 1996 (dashed-dot line) and 1997 (dashed line). Also shown is the approximate 2-week average evapotranspiration. 'Day' on the x -axis is the center point of the 2-week average.

emergence of new leaves. Soil moisture was at or near field capacity before and after the spring transition each year (Fig. 2). Similar rapid and extreme changes in Bowen ratio have also been observed due to leaf emergence in a southern boreal aspen stand (Blanken et al., 1997) and a deciduous forest in Massachusetts (Moore et al., 1996). The transition was marked, but less dramatic, over a Japanese grassland (Saigusa et al., 1998). Leaf emergence occurred latest in 1996 (Fig. 3, Table 1), but this difference is detectable only as a slight delay in the initial increase in latent heat fluxes in the two 2-week averages following day 100 (Fig. 5).

Latent heat flux decreased more during the middle and late summer in 1995 and 1996 than in 1997. During the stress period in 1995 the Bowen ratio during the daytime was about twice those in 1996 and 1997. By day 300 sensible heat flux was again larger than latent heat flux for all 3 years. This autumnal transition in fluxes was less intense and more temporally diffuse than the spring transition, likely because the senescent process was less temporally distinct and the atmosphere was warmer in late summer and early fall, suppressing sensible heat flux. Daily sensible heat fluxes were also reduced by relatively large negative

fluxes from the warm soil at night during this period. The late season transition in fluxes was similar at Harvard forest (Moore et al., 1996). A large transition in fluxes occurred during senescence in a North American grassland (Ham and Knapp, 1998), while the effect of senescence in a Japanese grassland was also less distinct (Saigusa et al., 1998).

These variations in the Bowen ratio, due to soil moisture deficits or senescence and most dramatically to leaf expansion, have significant impacts on properties of the atmospheric boundary layer. 'Green-up' decreases sensible heat flux and likely reduces boundary layer turbulence and boundary layer depth, evidence that phenology may be a controlling variable in dispersion modeling, convective storms and hydrology. For a given atmospheric profile, the boundary layer dries and warms most rapidly in later winter and early spring and this warming and drying likely decreases just after leaf emergence. This phenomenon appears to be detectable at regional scales; at a number of stations leaf emergence often coincides with increased humidity and decreases in the diurnal temperature range (Schwartz, 1996), as well as increased cumulus development in the boundary layer (Freedman et al., 1998).

3.8. Surface conductance

The distribution of sensible and latent heat flux, and the resulting effects on boundary layer processes, is dependent on both biological and climatological controlling variables. Fig. 5 shows that the emergence of leaves acts as a switch that transfers energy from sensible to latent heat flux. Other biological or climate controls within and between years also affect this energy distribution, and in the following sections we attempt to estimate the magnitude of these controls.

‘Biological’ control will be defined by the effect of changes in surface conductance on evapotranspiration. Surface conductance, calculated from (1), does not explicitly represent the physiological parameter stomatal conductance because it can include nonlinear effects of soil moisture and canopy turbulence (Rau-pach and Finnigan, 1988; Paw and Meyers, 1989). Despite these difficulties, theoretical and experimental investigations indicate that surface conductance is often related to the weighted integration from individual leaves (Granier and Loustau, 1994; Herbst, 1995; Lu et al., 1995; Raupach, 1995; Baldocchi and Meyers, 1998). During the dormant season, soil evaporation determines the surface conductance. Soil evaporation can also affect surface conductance during the growing season, but soil evaporation at this forest is typically small ($25\text{--}30\text{ W m}^{-2}$) (Baldocchi and Meyers, 1991), and it is assumed that this contribution to surface conductance is relatively minor.

Average surface conductance was calculated for each hour from (1) and for each day and 2-week period from (2). Conductances were only calculated on days with no precipitation (dry surface conductance). In this analysis we assume that during the growing season surface conductance responds primarily to two different biological phenomena: changes in leaf area and changes in leaf-level stomatal conductance. The latter change can occur through changes in photosynthetic capacity or through humidity deficits, soil water content and solar radiation. During leaf expansion, surface conductance changes primarily from increasing leaf area. Fig. 6 shows that changes in leaf area and surface conductance were correlated during this transition period, although photosynthetic capacity of current leaf area was also possibly increasing. Similar relationships have been found in an oak forest (Granier and Breda, 1996), an aspen forest (Blanken

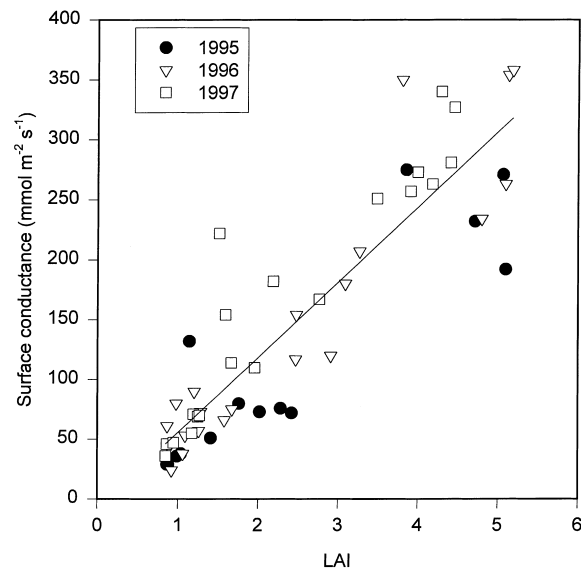


Fig. 6. Daily surface conductance during leaf expansion for all 3 years as a function of leaf area index. The line is the least-squares regression (intercept = $-7.0\text{ mmol m}^{-2}\text{ s}^{-1}$, slope = $62.6\text{ mmol m}^{-2}\text{ s}^{-1}$, $R^2 = 0.81$).

et al., 1997) and a Japanese grassland (Saigusa et al., 1998).

Typical hourly maximum dry surface conductances over the 3-year period were about $700\text{ mmol m}^{-2}\text{ s}^{-1}$, which is within typical values found for deciduous and coniferous forests (Kelliher et al., 1995). For the mid-summer period when leaf area is nearly constant, changes in surface conductance are assumed to largely reflect changes in stomatal conductance. Fig. 7 shows the 2-week averages of surface conductance for the 3 years calculated from (2). Typical 2-week averages of surface conductance for most of the growing season were around $400\text{ to }500\text{ mmol m}^{-2}\text{ s}^{-1}$ in 1996 and 1997. In the drought year (1995) surface conductance was only about one-half or less these values from July through the remainder of the season.

After the rapid increase in surface conductance with leaf emergence, surface conductance either reached a plateau (1996), decreased moderately (1997) or decreased dramatically (1995) for the remainder of the season. These different trends for each of the 3 years are correlated with changes in soil water content over the season (Fig. 2). Granier and Breda (1996) present a similar pattern in a French oak forest where surface conductance was a function of leaf area when

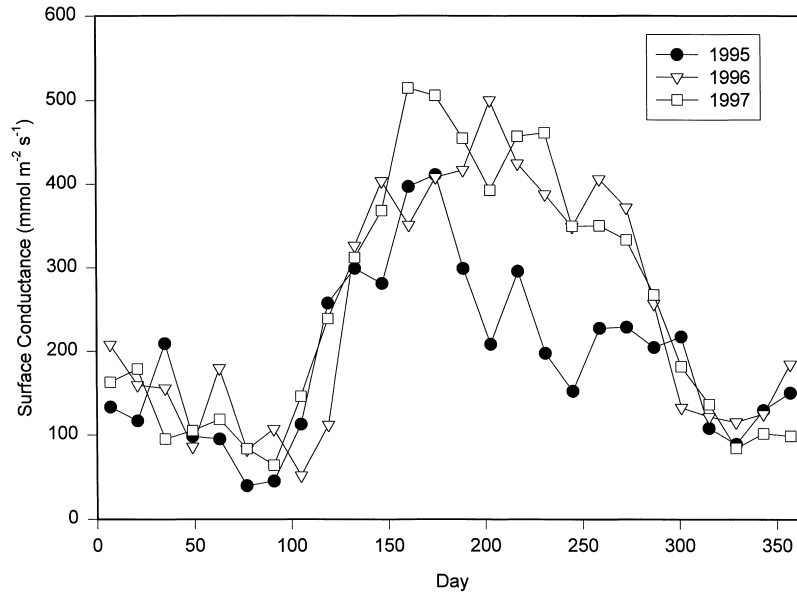


Fig. 7. Two-week averages of the surface conductance for each of the 3 years. 'Day' on the x -axis is the center point of the two-week average.

soil moisture was plentiful but soil water content controlled conductance as the soil dried.

3.9. Biological control of evapotranspiration

Likely due to stomatal closure, surface conductance was significantly less during the summer of 1995 compared to the other years, which tended to decrease evapotranspiration. Alternatively, the humidity deficit was also greater in 1995, which alone should increase evapotranspiration. Although it is not possible to completely separate biotic and abiotic influences, this shows the presence of annual variability in both the 'biological' and climatological parameters that drive evapotranspiration. A significant hydrological and meteorological question is the relative importance of biological and climate parameters in creating seasonal and interannual variability. If atmospheric demand is the primary control of variability, then evapotranspiration between years could be estimated directly from radiation and humidity deficits, while changes in phenology, drought physiology and stomatal responses could be treated peripherally. Alternatively, if biological responses are needed to adequately estimate annual evapotranspiration, then

accurate representations of phenology and stomatal response are necessary.

Control by atmospheric demand will be defined as the effect of net radiation (R_n) and ambient humidity deficit above the canopy (D) on evapotranspiration between years. The net radiation and humidity deficit contributions will be evaluated separately. Biological control will be assessed from changes in evapotranspiration due to surface conductance. In this analysis, the Penman–Monteith equation is written (Jarvis and McNaughton, 1986):

$$E = \Omega E_{eq} + (1 - \Omega) \rho G_c D + \Omega Ib \quad (3)$$

where $LE_{eq} = \varepsilon/(\varepsilon + 1) R_n$ is the equilibrium evaporation. The imbalance term $Ib = [\varepsilon/(\varepsilon + 1) (R_n - G - C_s - H - LE)]$ is not normally part of the Penman–Monteith equation, but arises because the calculation of surface conductance in (1) is based on the measured Bowen ratio and not net radiation. More simply, it represents the measurement error due to non-closure of the energy balance. Ω is the 'decoupling coefficient':

$$\Omega = \frac{1 + \varepsilon}{1 + \varepsilon + G_a/G_c} \quad (4)$$

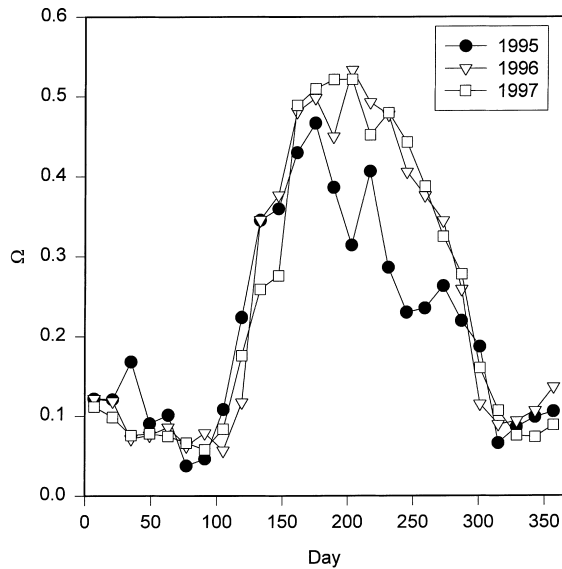


Fig. 8. Two-week average of the decoupling coefficient for each of the 3 years. 'Day' on the x-axis is the center point of the 2-week average.

Average 2-week values for the decoupling coefficient on days with no precipitation are shown in Fig. 8. Small values of Ω , typical for coniferous forests, indicate that evapotranspiration is highly sensitive to surface conductance and ambient humidity deficit. High values of Ω , typical of agricultural species, indicate that evapotranspiration is more sensitive to net radiation. Average mid-summer values of Ω are around 0.5 in 1996 and 1997, suggesting that the sensitivity of evapotranspiration to relative changes in humidity deficit, stomatal conductance and net radiation are similar. This forest is less sensitive to surface conductance than some other broadleaved forests (Meinzer, 1993; Herbst, 1995; Granier and Breda, 1996) and especially coniferous forests (Kelliher et al., 1993; Meinzer, 1993). Values of Ω for tropical forests are greater than those in other forest types, including the forest in our study (Meinzer, 1993). In 1995, Ω was smaller and surface conductance and humidity deficit exerted greater control over evapotranspiration. For the moment we assume Ω and ρ are constant, and the total derivative of (3) is:

$$dE = \frac{\partial E}{\partial G_c} dG_c + \frac{\partial E}{\partial D} dD + \frac{\partial E}{\partial R_n} dR_n + \frac{\partial E}{\partial Ib} dIb \quad (5)$$

An approximate form of (5) can be used to estimate the

cause of variability in E (ΔE) between two different years:

$$\Delta E = \frac{\overline{\partial E}}{\partial G_c} \Delta G_c + \frac{\overline{\partial E}}{\partial D} \Delta D + \frac{\overline{\partial E}}{\partial R_n} \Delta R_n + \frac{\partial E}{\partial Ib} \Delta Ib \quad (6)$$

where Δ represents the difference in E , G_c , D , R_n and Ib between two different years and the overbar represents the mean between the 2 years. The partial derivatives and differences were evaluated for each 2-week period during the seasons using the 2-week averages of G_c , G_a , R_n , D , ε and Ib . For each 2-week period of each year the four terms on the right-hand side of (6) were obtained from the partial derivatives of (3).

Since 1997 had the highest rates of evapotranspiration, (6) was used to evaluate the processes [three terms on right on side of (6)] or measurement error [fourth term in (6)] that reduced evapotranspiration in the drought year of 1995. Although this analysis is clearly an approximation, for each two week period the four terms on the right-hand side of (6) were very similar to the measured differences in evapotranspiration (ΔE).

The first term on the right-hand side represents the difference in evapotranspiration between the two seasons due to changes in surface conductance. The second term is the difference in evapotranspiration due to differences in the humidity deficit, while the third term is the difference in evapotranspiration due to changes in net radiation. The last term represents an error due to lack of energy balance closure in the measurements. Note that each of the four terms in (6) is the product of a sensitivity and a difference. While the relative sensitivity of evapotranspiration to G_c , D and R_n were fairly similar for much of the growing season (Ω was often near 0.5), the actual difference between years also depends on the how these variables change between years.

The magnitude of 'atmospheric demand' (humidity deficit, net radiation), 'biological' (surface conductance) and 'error' (imbalance) parameters in reducing the measured dry-canopy evapotranspiration for the entire years of 1995 and 1996 compared to 1997 are presented in Fig. 9. For all the measurement hours in this analysis (net radiation exceeds zero and no precipitation during day), evapotranspiration was reduced

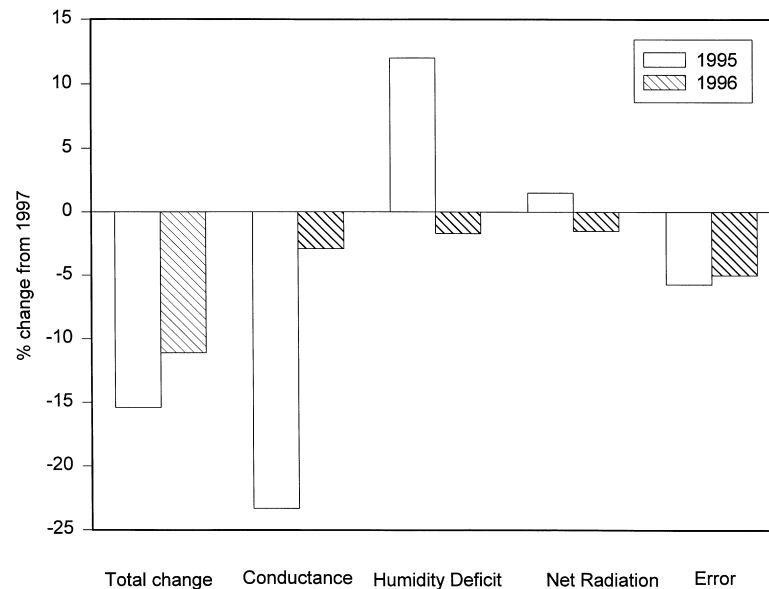


Fig. 9. The percent change in dry canopy evapotranspiration rate (mm h^{-1}) in 1995 and 1996 compared to 1997 (bar labeled 'Total change'). Also shown is the percent change in evapotranspiration rate (mm h^{-1}) derived analytically from (6) due to changes in surface conductance, humidity deficit, net radiation and energy balance closure errors.

by 15.4% in 1995 and by 9.9% in 1996 compared to 1997 (Fig. 9). Fig. 10 shows the magnitude of the four terms evaluated on a 2-week basis between 1995 and 1997. For the entire season, the reduction in evapotranspiration during 1995 compared to 1997 was primarily due to a reduced surface conductance, especially during the middle and late summer (Figs. 9 and 10). If the only difference between the seasons had been the surface conductance, dry-canopy evapotranspiration would have been reduced by approximately 23.3% in 1995 to that in 1997 (Fig. 9). With this interpretation, we can approximate a 23% decrease in evapotranspiration in 1995 to 'biological' control.

Counteracting this effect, the humidity deficit was larger in 1995 than in 1997, especially during the mid-summer period when the surface conductance was smaller (Fig. 10). If the only difference between the two years had been the increase in humidity deficit, evapotranspiration would have been about 12.5% *greater* in 1995 than in 1997 (Fig. 9). Therefore, controls by surface conductance and humidity deficit strongly opposed each other, consistent with model simulations performed over the 1995 season (Baldocchi, 1997). Net radiation was greater in 1995

and this affect is estimated to have increased evapotranspiration in 1995 by 1.5% compared to 1997 (Figs. 9 and 10). The error term due to the energy imbalance was responsible for a 5.6% decrease in the measured evapotranspiration in 1995 relative to 1997 (Figs. 9 and 10).

This analysis oversimplifies the complexity and the potential feedbacks involved between surface conductance and humidity deficit (McNaughton and Jarvis, 1991; Jacobs and De Bruin, 1991; Davis et al., 1997; Jacobs and De Bruin, 1997). The humidity deficit of the mixed layer responds to surface conductance because the surface conductance modifies the heat and humidity fluxes (Jacobs and De Bruin, 1991). If we assume that the higher humidity deficit in 1995 was exclusively due to a feedback from the lower surface conductance, then it is not accurate that 'biological control' reduced evapotranspiration by 23%. In this case the larger humidity deficit, which theoretically would have increased evapotranspiration by roughly 11.5%, was a feedback from the lower surface conductance. The higher humidity deficit would not have occurred without the change in surface conductance, so the total 'biological control' could be roughly esti-

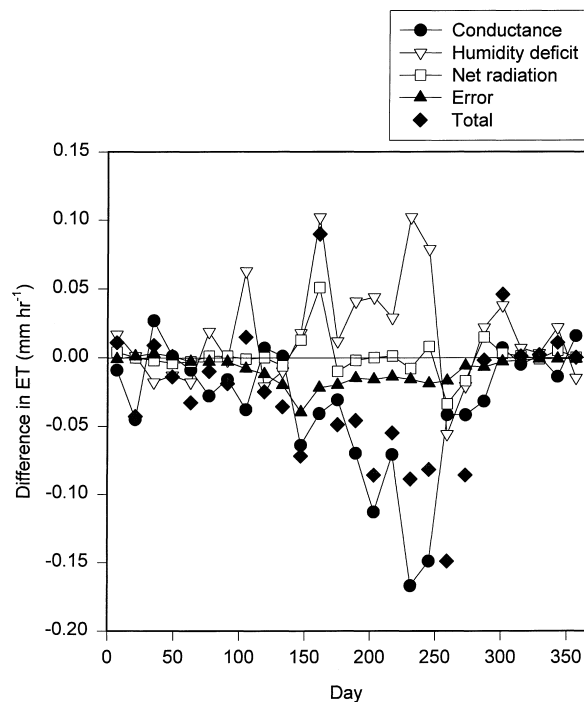


Fig. 10. The four terms on the right-hand side of Eq. (6) evaluated for each two-week period between 1995 and 1997 when the canopy was dry. The total measured difference in the average rate of evapotranspiration between the 2 years is also shown. 'Day' on the x-axis is the center point of the 2-week average.

mated to be $23.3\% - 11.5\% = 11.8\%$. Alternatively, if the humidity deficit is determined independent of surface conductance (e.g. synoptic scale advection) then biological control is roughly 23%. In reality the extent of feedback on the mixed layer is likely between the two scenarios presented here, but these two extremes set reasonable bounds on the extent that surface conductance reduced dry-canopy evapotranspiration in 1995 relative to 1997. As suggested from this analysis, biological control decreases when the humidity deficit is determined more by regional scale feedbacks (McNaughton and Jarvis, 1991; Jacobs and De Bruin, 1991) than by advection (McNaughton, 1976; Shuttleworth and Calder, 1979).

3.10. Priestly–Taylor Coefficient

Evapotranspiration is often normalized to the equilibrium value in (3) ($\alpha = E/E_{eq}$) as a way of comparing measured evaporation to a climatological expectation

assuming a closed volume and constant net radiation over a 'wet' surface (McNaughton and Jarvis, 1983). It could then be suggested that deviations from 'wet' surface evaporation are attributable to limiting physiological constraints (stomatal conductance) or to advective processes that prevents equilibrium conditions. Evapotranspiration rates greater than equilibrium ($\alpha > 1$) can occur through horizontal advection (Singh and Tallefer, 1986) or mixed layer entrainment (Culf, 1994). Although there is no basis for expecting α to converge to any theoretical limit, measurements and modeling studies indicate that α is commonly somewhat greater than one (between 1.1–1.4) for well-watered crops or 'wet' surfaces (Lindroth, 1985; McNaughton and Spriggs, 1989; Lhomme, 1997; Kim and Entekhabi, 1997). At our site, typical maximum growing season values of α during days with rainfall (wet canopy) were around 1.1, rarely exceeding 1.3.

Fig. 11 shows the daily daytime ($R_n > 0$) value of α against dry surface conductance for the three

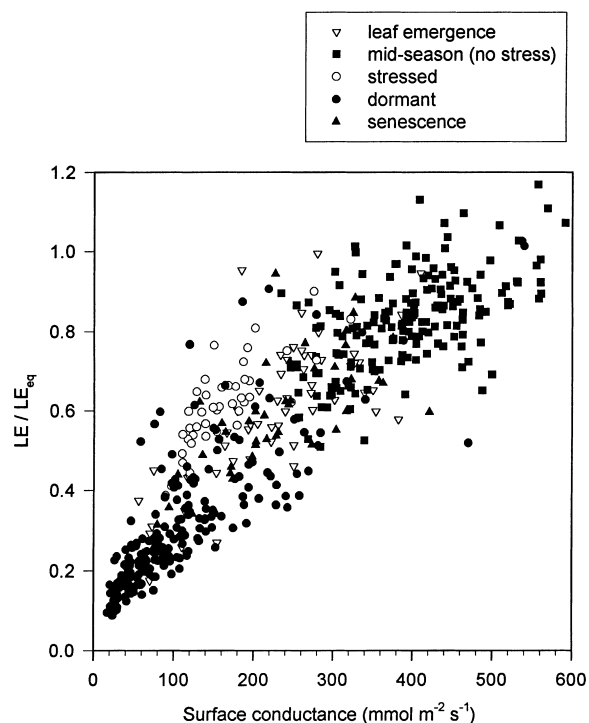


Fig. 11. The average daily value of α (LE/LE_{eq}) as a function of the daily surface conductance during different phenological stages or periods of stress. Data from all years are combined, except stressed values are from the mid-season of 1995.

seasons. The symbols represent four different phenological stages of the canopy (dormant, leaf expansion occurring, mid-season, senescent) along with data from the stress period during 1995. Variability of α , or deviations from the idealized 'wet surface', are created largely by changes in surface conductance, which we have shown to be related to changes in leaf area and stomatal conductance. Model sensitivity studies also show that α is a function of the product of photosynthetic capacity and leaf area index until a threshold of this product is reached (Baldocchi and Meyers, 1998). During the FIFE study, both spatial and temporal variability in α were a function of leaf area and soil water content (Chen and Brutsaert, 1995).

The daily value of α rarely exceeds one at any time of the year, suggesting that the equilibrium value approximately establishes an upper limit to dry canopy evapotranspiration for this forest. In the model simulations of McNaughton and Spriggs (1989) and Kim and Entekhabi (1997), α became insensitive to surface conductance when surface conductance exceeded about $700 \text{ mmol m}^{-2} \text{ s}^{-1}$. Maximum daily surface conductances at our site are just below those thresholds, and surface conductance appears to limit α during all phenological stages (Fig. 11).

Using α as a tool for comparison between sites has the advantage of normalizing sites against equilibrium rates determined primarily by net radiation. The 'daytime mean' α was 0.91 during one growing season at the aspen boreal site (Blanken et al., 1997), which is greater than our three-year growing season 'daytime mean' value (0.70) (Table 1). The twenty four-hour α at a tropical rainforest was 0.91 for 9-day period (Shuttleworth et al., 1984), greater than the 24 h value at our forest (0.80) during the growing season (Table 1). Since the error in energy balance closure at our site could account for these small differences it is not certain to what extent these studies actually contrast with ours.

4. Summary and conclusion

Based on 3 years of analysis, interannual variability of evapotranspiration at this site was about 10–15% of the 3-year mean (567 mm). Over the study period, the latent heat flux was about 30% greater than sensible heat (annual Bowen ratio averaged 0.72), but the rel-

ative partitioning at any given time was strongly dependent on whether leaves were present or not. Evapotranspiration at this site was still typically less than equilibrium during all periods, except when the canopy was wet. Interannual variability was determined primarily by drought and corresponding annual differences in surface conductance, with smaller influences by annual differences in net radiation and phenology. However, the effect of drought on evapotranspiration was probably decreased significantly due to feedbacks on the mixed layer humidity deficit.

Characterizations of bulk parameters at specific sites have potential benefits for identifying and quantifying exchange processes across the globe. Because of the complexity in representing canopy architecture, turbulence and physiology for the entire globe, bulk parameters are useful characterizations of exchange processes (Sellers et al., 1992; Jarvis, 1995; Kelliher et al., 1995). As longer term data sets become available it should be possible to examine whether these parameters are functions of other more fundamental scaling parameters within the ecosystem such as leaf area, soil moisture, leaf nitrogen, length of growing season and precipitation ratio (precipitation/equilibrium evaporation) (Baldocchi and Meyers, 1998). The best approach will be combining analysis of continuous data sets along with canopy exchange models and physiological and boundary layer studies to demonstrate the links between these processes.

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