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Partitioning evapotranspiration using stable isotopes and Lagrangian dispersion analysis in a small agricultural catchment

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Abstract: Measuring evaporation and transpiration at the field scale is complicated due to the heterogeneity of the environment, with point measurements requiring upscaling and field measurements such as eddy covariance measuring only the evapotranspiration. During the summer of 2014 an eddy covariance device was used to measure the evapotranspiration of a growing maize field at the HOAL catchment. The stable isotope technique and a Lagrangian near field theory (LNF) were then utilized to partition the evapotranspiration into evaporation and transpiration, using the concentration and isotopic ratio of water vapour within the canopy. The stable isotope estimates of the daily averages of the fraction of evapotranspiration (Ft) ranged from 43.0–88.5%, with an average value of 67.5%, while with the LNF method, Ft was found to range from 52.3–91.5% with an average value of 73.5%. Two different parameterizations for the turbulent statistics were used, with both giving similar R^2 values, 0.65 and 0.63 for the Raupach and Leuning parameterizations, with the Raupach version performing slightly better. The stable isotope method demonstrated itself to be a more robust method, returning larger amounts of useable data, however this is limited by the requirement of much more additional data.

Keywords: Evapotranspiration partitioning; Stable isotopes; Lagrangian dispersion theory.

INTRODUCTION

Evapotranspiration (ET) forms an important part of the water balance across all spatial and temporal scales. For many applications however, further information on the partitioning of evapotranspiration into its constituent components of evaporation and transpiration is required, e.g. for irrigation management (Tong et al., 2009) and climate modelling (Lian et al., 2018).

At different spatial scales ET can be estimated using different methods and techniques such as water balance, remote sensing and energy balance modelling at global and regional scales (Vinukollu et al., 2011), or measured using eddy covariance or scintillometry at field scales. However, these methods provide little or no information on how the ET is partitioned into its components of evaporation (E) and transpiration (T). Other direct measurement techniques at the point scale, such as lysimeters for measuring soil evaporation (Heinlein et al., 2017; Rafi et al., 2019) or sap flow sensors for transpiration (Agam et al., 2012; Zhao et al., 2016) can be used to directly measure the individual components, however they have a very small footprint which requires upscaling to the field scale, which is strongly influenced by the heterogeneity of the surrounding area. For trees in the riparian zone, transpiration can be estimated from groundwater fluctuations (Gribovszki et al., 2008), however this is not applicable for crops with their shallower root systems. Assumptions can be made in order to estimate Ft, such as when E or T could be assumed to be negligible, however these assumptions have been shown to not be applicable to every ecosystem (Stoy et al., 2019). An alternative approach is to partition the measured field scale ET using methods such as Lagrangian dispersion analysis (Raupach, 1989a; Warland and Thurtell, 2000) or stable isotopes (Ma et al., 2018; Wang et al., 2013; Williams et al., 2004; Xiao et al., 2018; Yepez et al., 2003).

The Lagrangian dispersion analysis method an inverse modelling approach, where the source/sink distribution of a scalar is related to a concentration profile of the scalar within a canopy ecosystem through a turbulent dispersion field (Raupach 1989a; Santos et al., 2011; Warland and Thurtell, 2000). Within a plant canopy, the gradient - diffusion relationship cannot be used to calculate the flux of a trace gas due to the role of large turbulent eddies in the vertical transport of the scalar, resulting in the observation of counter-gradient fluxes (Denmead and Bradley, 1985). Instead Raupach (1989a, 1989b) developed a model in a Lagrangian framework where the trajectory of fluid particles released from a source is followed, thereby taking into account the history of the particles. Then the resulting concentration profile can be broken up into two regions, a near field region where the particles disperse linearly in time due to the persistence of the turbulent eddies, and a far field region where the particles disperse with the square root of time diffusively. The method has been used to calculate the flux of CO_2, H_2O and heat within coniferous (Styles et al., 2002) and eucalyptus forest canopies (Haverd et al., 2011). In addition, Haverd et al. (2009) applied the method in a eucalyptus forest along with a Soil-Vegetation-Atmospheric-Transfer model (SVAT) to improve the estimation of turbulent statistics in a forest canopy.

The Keeling Plot (Keeling, 1958) mass balance can be applied within a field ecosystem, to the isotopes of water evaporated from within the canopy. As water evaporated from the soil has a different ratio of light to heavy isotopes compared to water vapour evaporated from plant stomata, this difference in isotopic ratio allows for the estimation of the soil evaporation-

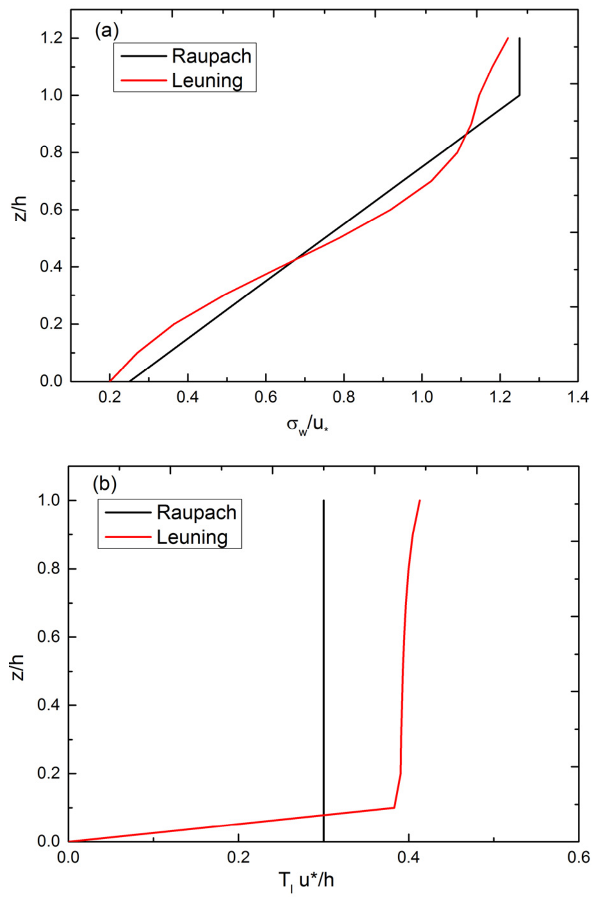

Fig. 1. Normalised profiles of the standard deviation of (a) the vertical velocity, σ_w and (b) the Lagrangian time scale, T_L, according to Raupach (1989a) and Leuning (2000).

transpiration ratio within the ecosystem. This method has been tested in a number of different vegetation and crop types such as an olive orchard (Williams et al., 2004), woodland (Dubbert et al., 2013), beets (Quade et al., 2019) and maize (Wu et al., 2016). Applying the method to determine the temporal variability of root water uptake depth of wheat plants, Zhang et al. (2011) estimated that up to 30% of water consumption during the irrigation season can be evaporation. Using deuterium isotopes in grassland over a short time period, Good et al. (2014) investigated the evolution of the evaporation/transpiration ratio during the growing phase of the grass crop. Wei et al. (2015) measured Ft over a rice field over the course of an entire growing season, with the estimated Ft ranging from 0.2 at the start of the growing season up to a near constant value between 0.8–1.0.

The objective of this study is to estimate the components of evapotranspiration in a vegetative maize field and its response to precipitation events at a high temporal resolution. In this paper we use the stable isotope method and the localised near field theory of Lagrangian dispersion analysis. Both methods use measurements of water vapour concentration within the canopy and a mass balance approach as their basis, while the isotope method requires extra measurements and assumptions, this allows for an additional study of the robustness and effectiveness of the two methods.

THEORY AND METHODS

Lagrangian dispersion analysis

In order to partition the evapotranspiration inside a crop ecosystem between E and T using an inverse method we first divide the source distribution into m vertical layers, where the first layer is just above the surface to account for soil evaporation, and the subsequent layers extend from just above this layer to the top of the canopy for transpiration. The water vapour released from these layers results in a concentration profile, which can be measured at n different heights. The relationship between scalar source density $\varphi(z)$ and concentration $C(z)$ under steady conditions in a horizontally homogenous canopy can be written as

$$C_i - C_r = \sum_{j=1}^{m} D_{ij}\varphi_j \Delta z_j \tag{1}$$

where C_i is the concentration at height z_i, C_r is the concentration at a reference height above the canopy z_r, D_{ij} is the dispersion matrix, with i rows for 1, .., n measurement heights, and j columns for 1, .., m source layers , φ_j is the source strength in layer j, and Δz_j is the thickness of layer j. The inversion of this equation then allows for the estimation of the source strengths, in this case E and T. To calculate the dispersion matrix, the source strength in one layer j is set to be a steady unit source, S_j and set to zero in all other layers. This gives a partial concentration profile C_i, which defines the elements of D_{ij} for dispersion from layer j to the concentration at height z_i

$$D_{ij} = \frac{C_i - C_r}{S_j - \Delta z_j} \tag{2}$$

In a Lagrangian analysis of particle dispersion within a canopy, the trajectory of a particle is followed from its release from a source. The resulting concentration C_i is due to the influence of two different regions on this path, a near field and a far field region.

In the Localised Near Field (LNF) theory of Raupach (1989a, 1989b), the two regions are treated separately, with C_i equal to the sum of both regions

$$C_i = C_n + C_f \tag{3}$$

In the near field, local effects dominate the dispersion of the particles while in the far field, it is assumed that particles diffuse in accordance with gradient-diffusion theory. The near and far field components can be described using the turbulent statistics for the canopy, the vertical profile of the standard deviation of the vertical velocity (σ_w) and the Lagrangian time scale (T_L) according to Raupach (1989a) as

$$C_n(z) = \int_0^{\infty}\!\!\int \frac{S(z_s)}{\sigma_w(z_s)}\left\{k_n\left[\frac{z - z_S}{\sigma_w(z_s)T_L(z_s)}\right] + k_n\left[\frac{z + z_S}{\sigma_w(z_s)T_L(z_s)}\right]\right\} dz_S \tag{4}$$

$$C_f(z) = C(z_{ref}) - C_n(z_{ref}) + \int_z^{z_{ref}} \frac{F(z')}{K_f(z')} dz' \tag{5}$$

where z' is the discrete height, z_s is the source height, k_n is a near field 'kernel', $F(z)$ is the flux density, and K_f is the far field diffusivity.

$$k_n(\zeta) = -0.3989\ln\left(1 - \exp(-|\zeta|)\right) - 0.1562\exp(-|\zeta|) \tag{6}$$

$$F(z) = -K_f(z)\frac{dc_f}{dz} \tag{7}$$

$$K_f = \sigma_w^2(z) T_L(z) \tag{8}$$

where $\zeta = (z - z_0)/ (\sigma_{w0}\, T_{L0})$. As T_L cannot be directly measured by fixed sensors as it is a Lagrangian quantity and due to the difficulty of making vertical wind measurements inside a canopy, the profiles of σ_w and T_L needed for the calculation of D_{ij} are normally calculated using turbulent statistical parameterisations. Based on the results of Santos et al. (2011) the parameterisations suggested by Raupach (1989a) and Leuning (2000) were used in this study. Raupach (1989a) proposed that σ_w and T_L profiles could be approximated by the piecewise linear profiles

$$\frac{\sigma_w}{u^*} = f(x) = \begin{cases} a_0 + (a_1 - a_0) \times \dfrac{z}{h}, & z < h \\ a_1, & z \geq h \end{cases} \tag{9}$$

$$\frac{T_L u^*}{h} = \max\left[c_0, \frac{k(z-d)}{a_1 h} \right] \tag{10}$$

where a_1 = 1.25, a_0 = 0.25, c_0 = 0.3, d = 2/3h is the displacement height, h is the canopy height, u^* is the friction velocity and k = 0.41 is the von Karman constant.

The parameterisations of Leuning are based on exponential and non- rectangular functions within and above the canopy

$$y = c_1 e^{c_2 z/h} \quad \text{for} \quad z < 0.8h$$

$$y = \frac{(ax+b) + d_1\sqrt{(ax+b)^2 - 4\theta abx}}{2\theta} \quad \text{for} \quad z \geq 0.8h \tag{11}$$

where c_1 = 0.2 and the other coefficients can be found in Table 1.

Table 1. A list of the variables and parameters used for the determination of normalised profiles of the standard deviation of the vertical velocity, σ_w and the Lagrangian timescale, T_L. (Leuning, 2000).

z/h	x	y	θ	a	b	d
≥0.8	z/h	σ_w/u^*	0.98	0.850	1.25	–1
≥0.25	z/h – 0.8	$T_L u^*/h$	0.98	0.256	0.40	+1
<0.25	$4z/h$	$T_L u^*$/h	0.98	0.850	0.41	–1

These parameterisations were originally derived for near-neutral conditions and corrections functions have been suggested for use in non-neutral conditions (Leuning, 2000). However the performance of these corrections has been mixed, with Santos et al. (2011) reporting an increase in the overestimation of the latent heat flux when the corrections were used.

Stable isotope method

Measurements of isotopes are expressed as the ratio of heavy to light isotopes relative to the international standard and written in (δ) notation in per mil (‰).

The fraction of evapotranspiration that is due to transpiration can be calculated as (Yakir and Sternberg, 2000)

$$F_T(\%) = \frac{\delta_{ET} - \delta_E}{\delta_T - \delta_E} \tag{12}$$

where F_T(%) is (T/ET x 100), δ_{ET} is the isotopic composition of water vapor that has been evapotranspirated, δ_T is the isotopic composition of water vapor that has been transpired, and δ_E is the isotopic composition of soil water that has been evaporated. δ_T and δ_E can be estimated by vegetation and soil sampling whereas δ_{ET} is determined using an ecosystem mass balance equation,

$$\delta_{ebl} = C_a\left((\delta_a - \delta_{ET}) \left(\frac{1}{C_{ebl}} \right) + \delta_{ET} \right) \tag{13}$$

where δ_{ebl} is the isotopic composition of water vapor in the system boundary layer, C_a is the water vapor concentration in the atmosphere, δ_a is the isotopic composition of water vapor in the atmosphere and C_{ebl} is ecosystem boundary layer water vapor concentration. Using this linear relationship, measurements of the isotope ratio of the water vapor of the air at different heights within the canopy plotted versus the inverse of the concentration, will yield an estimate of δ_{ET} as the resulting y-axis intercept (Yakir and Sternberg, 2000).

δ_E is usually not measured directly, due to the difficulty of designing non-destructive sampling methods, instead it is indirectly calculated using the Craig-Gordon model and measurements of soil water at the evaporating front within the soil (Craig and Gordon, 1965). The Craig- Gordon model estimates the effects of fractionations on liquid water in the soil as it evaporates (Moreira et al., 1997)

$$R_E = \left(\frac{1}{\alpha_K} \right) \frac{\left(\dfrac{R_s}{\alpha^*} \right) - R_A h}{1 - rh} \tag{14}$$

where R_E is the molar ratio of heavy to light isotopes of: E, the evaporated water vapor, R_s, water in the soil, and R_A, air near the surface, rh is the relative humidity normalised by the saturation pressure at the surface, α_K is the kinetic fraction rate, taken to be 1.0189‰, Flanagan et al., 1991) and α^* is the equilibrium fractionation factor as a function of temperature (T) (Majoube, 1971).

$$\alpha^* = \frac{1.137 \times 10^6}{T^2} - \frac{0.4156 \times 10^3}{T} - 2.0667 \tag{15}$$

The isotopic signature of transpiration (δ_T) can be determined non-destructively using closed leaf vapor chambers or by measurement of stem water isotopic composition, assuming no isotopic fractionation in the transpiration process (Lin and Sternberg, 1993; Wang and Yakir, 2000).

STUDY AREA

The experiment was performed from the 24th June – 2nd July 2014 at the Hydrological Open Air Laboratory (HOAL) at Petzenkirchen, Austria (48°9’ N, 15°9’ E) (Blöschl et al., 2016). The catchment has an area of 66 ha, elevation ranges from 268–323 m above sea level, with a mean slope of 8%. Land use comprises of 87% agricultural crops, 5% grassland pasture, 6% forest and 2% paved surfaces. The local climate can be described as humid, with a mean annual precipitation of 823 mm/yr, with larger amounts of precipitation in summer than in winter. The mean annual temperature is 9.5°C. Evapotranspiration in the years from 2013–2017 ranged from 442–518 mm/yr. As the experiment was planned for early summer and a limited time period, a 4.8 ha maize field was selected as the early growing stage and wider spacing between crops would allow for assumptions of turbulent mixing within

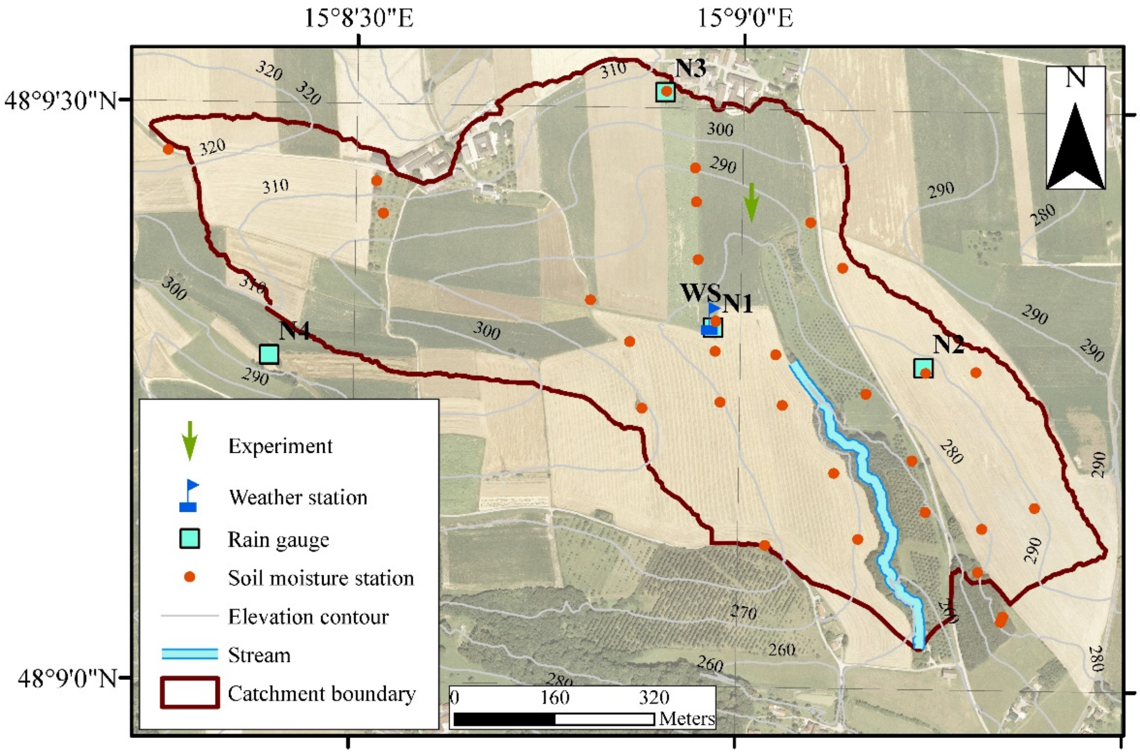

Fig. 2. The experimental catchment showing the location of the measurement devices and the study field (green arrow).

Fig. 3. The experimental area, showing the location of the eddy covariance system (right) and the picaro device (left).

the ecosystem to be fulfilled. The average height of the plants increased from 0.95 m to 1.40 m with the Leaf Area Index (LAI) progressing from 1.2 to 2.4 during this period.

Instrumentation

To measure the profiles of water vapour concentration and isotopic ratio (O_{18}/O_{16}) within the canopy a L2130-i analyser (Picarro) was installed within the maize field. Air was sampled within the canopy and above, using a 6-port intake valve connected to a pump and sampled using the analyser at 1 Hz. To achieve a precision of 0.02‰ an averaging time of at least 100 seconds was required for each individual ports and gave an overall resolution of 20 minutes for δ_{ET} and C across the 6 ports. δ_{ebl} and C_{ebl} were sampled at 4 heights within the canopy (0.1, 0.2, 0.5 and 1.0 m), and 2 additional sample intakes were located above the canopy (1.7 and 2.4 m). The ports at heights 0.5 and 1.0 m were later increased to 0.8 and 1.2 m on the 1st July due to the increase in canopy height. δ_T was estimated from xylem water taken from 4 maize plants, sampled between 11–14 h on each day of the experiment. This assumes that the xylem water is at isotopic steady state is normally is valid between late morning and early afternoon. To estimate δ_E using Equation (14), soil samples were taken daily at 4 locations near the air intake, at depths of 0–2, 2–5 and 5–10 cm to correctly identify the evaporation front according to Rothfuss et al. (2010). All soil and plant samples were sealed in glass vials, frozen and then transported to the laboratory for extraction, according to the guidelines of Mayr et al. (2016).

Evapotranspiration was measured using an open path eddy covariance sensor (IRGASON, Campbell Scientific). The device was installed in the middle of the maize field at a height of 2.20 m before the experiment and moved to a height of 2.80 m during the early morning of the 1st to stay above the minimum height limit described by (Aubinet et al., 2012). The TK3 software was used to calculate the latent heat flux from the raw measurements of the wind speed and water vapour (Mauder and Foken, 2015). As part of the processing procedure a number of corrections must be applied to the raw data: (i) a double rotation of the coordinate system, this was used rather than the planar fit method due to the rapid growth of the maize crop and short time period of the experiment, (ii) the Moore correction for high frequency loss (Moore, 1986), (iii)a sonic air temperature for the sensible heat flux, and (iv) the WPL correction to account for density fluctuations (Webb et al., 1980). The TK3 software includes a quality control analysis and sensible and latent heat flux data of a low quality were removed.

Air temperature and humidity were measured at the eddy covariance station using a HMP155 probe. Precipitation was measured across the catchment using 4 weighing balance gauges (OTT Pluvio). For this experiment the data from the closest rain gauge, located at the nearby weather station was used. The catchment is instrumented with a network of soil moisture stations utilising Time Domain Transmission (TDT) probes. A station was located in the maize field to measure the near surface soil temperature and water content at depths of 0.05 and 0.1 m.

RESULTS
Environmental conditions

Figure 4 shows the environmental conditions over the time period of the experiment. Weather conditions were mixed over the course of the experiment, with most days experiencing periods of sunshine and cloud, except for the 30th where the

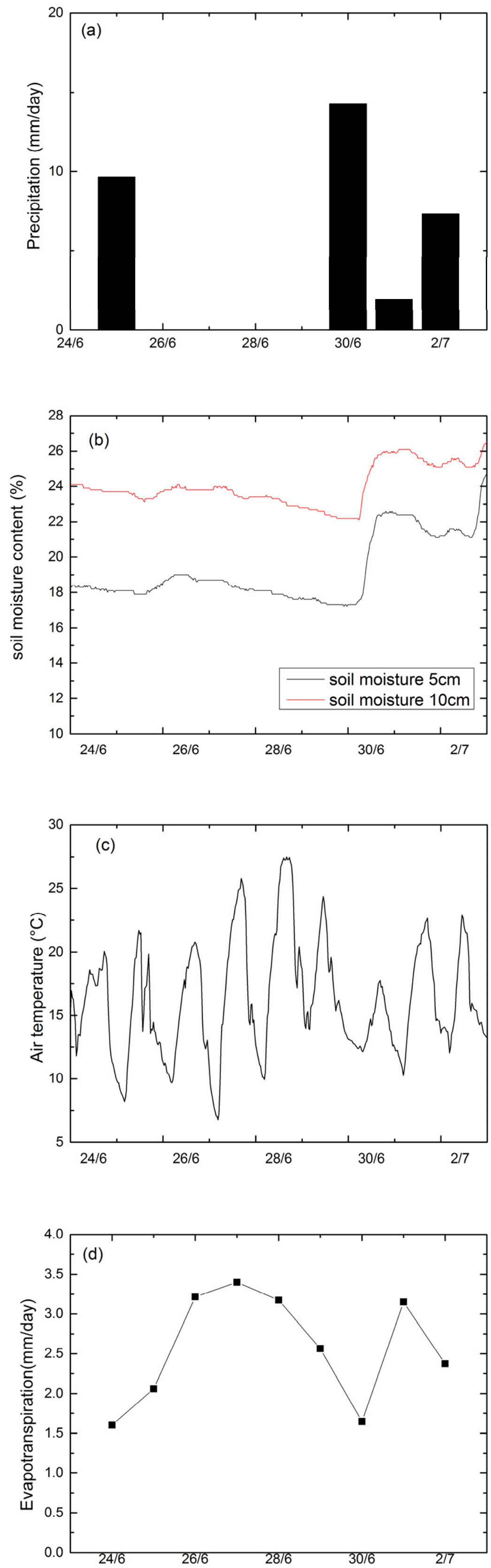

Fig. 4. Half hourly plots of (a) precipitation, (b) soil moisture at 5 and 10 cm, (c) air temperature and (d) daily values of evapotranspiration measured using the eddy covariance system for the period June 24th – July 2nd, 2014.

passage of a frontal system resulted in overcast conditions and persistent rain until late afternoon. In total four precipitation events were recorded, with three of them having a major effect on the near surface soil moisture level. The duration of the event on the 30th meant that it was not possible to use the data from this day, however the shorter nature of the other events meant less loss of data on those days. The average daily mean temperatures ranged from 14.2–20.0°C with maximum daily temperatures between 17.8–27.5°C. Daily evapotranspiration was strongly related to temperature, VPD and net radiation, with a total of 23.2 mm recorded over the experiment, with the highest daily values occurring during the dry period from the 26th – 28th. Figure 5 shows the friction velocity and the virtual stability measured using the eddy covariance system. The friction velocity was in general quite low at this site, following a diurnal pattern during the period of high solar radiation from the 26th – 29th. The atmospheric stability was generally unstable during the daytimes except for the period during the passage of the frontal system, where the stability was very close to neutral.

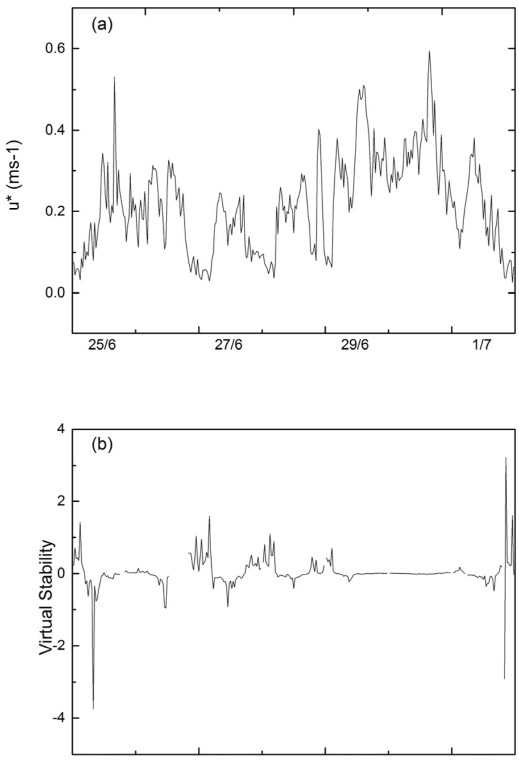

Fig. 5. Half hourly plots of (a) u* and (b) virtual stability over the period June 25th – July 2nd, 2014.

Evapotranspiration partitioning

The steady state assumption for the stable isotope analysis does not hold in the morning, however, it is usually met in the afternoon (Yepez et al., 2005). The analysis in this paper is hence limited to the time from 10:00–17:00, as this corresponds to the time periods with the largest amounts of solar radiation and hence evapotranspiration, there will only be a limited effect on the results. Data from the 30th and during and directly after precipitation events are also excluded, due to the strong neutral stability and change in the atmospheric water vapor resulting from the passage of the frontal system or interception, resulting in values of Ft in excess of 100%. However due to the very strong concentration gradients near the surface when soil evaporation is very close to zero, the inversion matrix can become ill-conditioned resulting in values over 100% for the LNF method. Therefore, in general the values of Ft higher than 110% and lower than –20% were excluded from the results (Wei et al., 2015). For purposes of averaging Ft was capped at 100%.

Using the stable isotope method, the daily averages of Ft_{ISO} ranged from 43.0–88.5%, with anaverage value of 67.5% following a pattern of decreasing after precipitation events and steadily increasing over the following days. Using the LNF method, Ft (Ft_{LNF}) was found to range from 52.3–91.5% with an average value of 73.5%. Figure 6 shows a comparison of the two methods for the entire experimental period. While the two methods show a high level of agreement on average, on the 29th and 1st July the LNF method estimates much higher values of Ft, on the 29th 91.5% versus 81.7% and on the 1st 52.3% versus 44.5%. Ft_{LNF} however shows much greater variance on these days than Ft_{ISO}, with Ft_{LNF} also estimating values of Ft over 100% on the 28th.

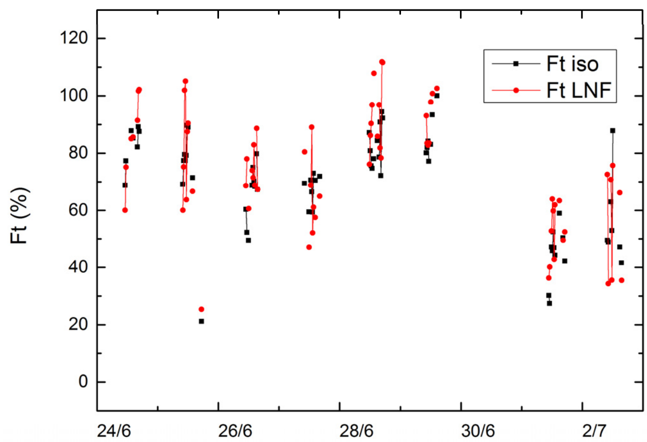

Fig. 6. Twenty-minute values of Ft using the LNF (red circles) and isotope (black squares) methods over the entire experimental period.

Daily ET is strongly dependent on solar radiation and temperature, with the highest values of ET measured from the 26th – afternoon of the 29th. Both Ft_{LNF} and Ft_{ISO} show a similar pattern, with increasing values estimated during this time period. Conversely the soil moisture content in the upper level of the soil at 5 cm was measured decreasing from 19.0% to 17.2%. Following the precipitation event on the 29th – 30th the soil moisture content was recharged up to 22.6%, leading to a marked decrease in Ft_{LNF} from 92.0% to 52.3%. Ft_{ISO} was found to be best correlated with solar radiation (R = 0.68) and showing less correlation with vapor pressure deficit (R = 0.59).

Method comparison

Figure 7 shows a comparison of the two methods on the 26th of June during the daytime period. The uncertainty on the Ft_{iso} estimates was calculated using the single isotope, two source mixing model of Phillips and Gregg (2001). In the late morning period, Ft_{LNF} consistently makes higher estimates of Ft, with a difference of up to 25.8% at 11:00, although closer agreement is noted during the afternoon period, following a gap in the results of Ft_{LNF} from 12:00–13:30 due to overestimation of Ft. During this period Ft_{ISO} continued to give realistic estimates of Ft.

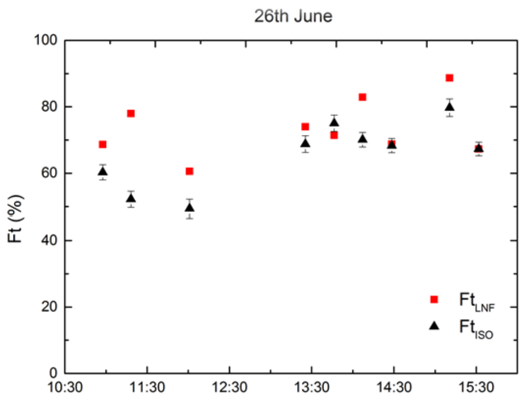

Fig. 7. Comparison of the two methods on the 26th of June during the daytime period.

Throughout the early afternoon Ft_{ISO} and Ft_{LNF} steadily increase with ET, with Ft_{LNF} increasing at a much faster rate to 80–90% by 15:00 while Ft_{ISO} exhibits a slower rate of increase, reaching a maximum of 80%.

Ft was also estimated using the Leuning set of parameterizations for $u*$ and T_L. In this case Ft_{LNF} varied from 50.1–91.4% with an average of 75.0%. Figure 8 shows a scatterplot of 20 minute values of Ft using the isotope and (a) LNF Raupach and (b) LNF Leuning methods. Both methods give similar R^2 values, 0.65 and 0.63 respectively with the Raupach parameterisations performing slightly better.

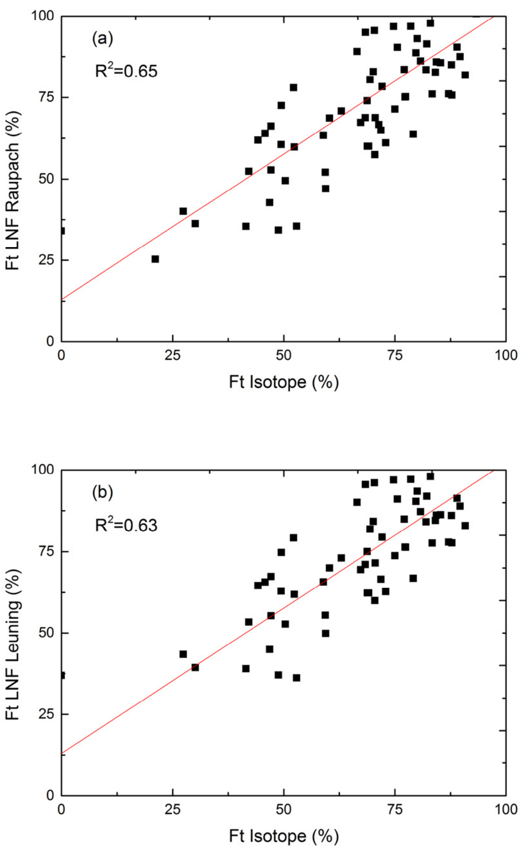

Fig. 8. Scatterplots of 20 minute values of Ft using the isotope and (a) LNF Raupach and (b) LNF Leuning methods for the entire experimental period.

DISCUSSION

In this study we utilize the LNF and ISO methods to partition evapotranspiration. Both methods use water vapor concentration measurements from inside the plant canopy, however, the stable isotope method requires much more additional data as well as equipment and labour for measuring and analysing the isotope samples.

The average Ft estimated by the LNF method was 74.0% and using the stable isotopes 67.5%. The correlation between the two methods was 0.65. These compare well with measurements using lysimeters which gave a range of 71–75% (Kang et al., 2003; Liu et al., 2002) for maize, 69–87% for olive trees (Williams et al., 2004), and 20–100% over the course of the entire season for rice (Wei et al., 2015). Using chamber based measurements to directly measure the isotopic values, Wu et al. (2016) estimated Ft for the entire vegetative growing section to be ~55%, and ~70% using the Craig- Gordon based model approach, with the difference attributed to deviation in measuring δE using the chamber method. Using isotope tracers Ma et al. (2018) found that for a winter wheat field the value of Ft over the entire crop season did not vary significantly with the type of irrigation treatment, however, the value of Ft for each stage of the growing season did. While an average value of for Ft 65% inline with similar experiments was found using 5 different irrigation methods, between these methods a difference of up 25% in Ft was noted. In non-irrigated catchments the precipitation amounts and intervals must be analyzed in order to apply the results from year to year. During the course of this experiment, both a short but intense (~10 mm/hr) and a less intense but longer duration (~1 mm/hr) precipitation event were recorded, allowing for the changes in Ft to be seen. The response of the soil moisture at 5 cm to the intense event was much less than for the longer event, with a large amount of runoff recorded. This results in a much smaller decrease in Ft on the following day, 73.4% versus 52.3%.

Over the course of the experiment the LNF method shows a pattern of slightly larger estimates of Ft, particularly on the 29th and 1st of July. On the 1st this exceptionally large difference is possibly due to the higher levels of soil evaporation after the large precipitation event on the previous day, resulting in reduced water vapor concentration gradients near the surface, which will have a greater effect on the LNF method as it uses less additional data. On the 29th Ft_{LNF} estimated Ft to be over 100%, however this can be explained due to measurement errors as Ft approached 100%, with Wei et al. (2015) reporting similar values while using the stable isotope method. The only slight change in the performance of the LNF method depending on the parameterisation for the turbulent statistics, would suggest either parameterisation can be used. Comparing LNF modelled and eddy covariance measured latent heat flux estimates, Santos ett al. (2011) also noted only slight changes to the results.

Over the course of the experiment the isotope method proved to be less affected by the environmental conditions, excepting the steady state condition that limits the method to the daytime periods. During these periods the LNF method gives 45.2% less data than the isotope method, with a lot less useable data on days where there is less coupling between the canopy and the atmosphere (25th and 29th) due to the changing conditions and precipitation. This reduction in results is offset however by the lower requirements of additional data, with the isotope method needing isotope measurements of the air, plants and soil. An advantage of using the LNF method which gives a high temporal resolution is that the response of the fraction of transpiration to rain events can be seen in Figures 6 and 8 and used for adjusting transpiration estimates measured using the eddy covariance method. Isotope methods that use only weekly sampling of the soil water (Santos et al., 2012) will not be able to capture changes in δ_E, however even the daily sampling of the soil and plant isotopes in this experiment is limited due to sudden short precipitation events and the isotopic steady state assumption. While the air sampling can be performed at a high resolution using the Picarro device, the sampling of the plants and soil had to be done manually in this experiment, resulting in a much higher workload and limiting the overall length of the measurement campaign. Measurement devices such as leaf and soil flux chambers which allow for automatic sampling of soil and plant isotope values have been developed in recent years (Wu et al., 2016), however they are still limited by the heterogeneity of field conditions, requiring a number of different devices of considerable expense.

With an estimate for Ft during the growing season, where ET and Ft change in response to not only the environmental conditions but due to the changing physical properties of the plant, compared to the more stable initial and mid-season stages of plant development (FAO-56). As it is difficult and expensive to make full season measurements at a high temporal resolution, an alternative approach is to use our estimate for Ft and an evapotranspiration model. Using a modified version of the FAO-56 method, where the crop coefficient is separated into a crop basal coefficient and an evaporation coefficient, Ding et al. (2013) was able to partition the ET by modifying the crop basal coefficient according to crop leaf cover. The model was then validated using soil heat flux and lysimeter measurements. During the vegetative season however the heterogenous canopy cover and rapid growth of the maize plants can lead to errors when upscaling the individual measurements to the field scale, to avoid this the LNF method which is based on vapor measurements allowing for an averaging through the canopy could be used.

CONCLUSIONS

In this experiment the fraction of evapotranspiration that was due to transpiration was estimated for a maize field using two methods, the isotope measurement based stable isotope method, and the Localised Near Field theory of Raupach based on an inverse Lagrangian modelling approach. Both methods are based on measurements of the water vapour concentration within a plant canopy, however they vary greatly in method. The two methods overall gave similar results, with the fraction of transpiration ranging from 43.0–88.5%, with an average of 67.5% for the isotope method, while the fraction of evaporation was found to range from 52.3–91.9% with an average value of 74.8% for the Localised Near Field method. These values were found to be in line with results from similar experiments for this stage of maize development. However, the stable isotope method was found to return a much larger amount of useable data, as well as having a lower variance. This is offset by the need for more additional measurements and analysis, as well the uncertainty due to the need for the Isotopic Steady State assumption. Future experiments should be conducted using chamber methods when possible to account for this. The parameterizations used for the turbulent wind statistics for the Localised Near Field method were found to vary only slightly. Care must also be taken when applying the results over larger time periods or year to year, to account for different precipitation regimes if the field is not irrigated, with different precipitation events giving different soil moisture and hence soil evaporation responses.

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