available at www.sciencedirect.com
ScienceDirect
journal homepage: www.elsevier.com/locate/agwat
ELSEVIER
Agricultural Water Management

Riparian zone equilibrium and actual evapotranspiration in a first order agricultural catchment in Southern Ontario, Canada

R.M. Petrone *, C. Smith, M.L. Macrae, M.C. English

Department of Geography & Environmental Studies, Wilfrid Laurier University, 75 University Avenue West, Waterloo, Ontario, Canada N2L 3C5

ARTICLE INFO

Article history:
Accepted 10 May 2006
Published on line 27 July 2006

Keywords:
Evapotranspiration
Priestley–Taylor coefficient
Riparian zone
Equilibrium evaporation
Lysimeter

ABSTRACT

Agricultural landcapes are typically comprised of a mosaic of land uses, some of which include riparian areas. While the active management of riparian water resources may not be the primary objective of many agricultural water managers, the importance of riparian zone evapotranspiration (ET) to over all basin water losses and water quality has been stressed in the literature. However, ET from these areas is rarely known with great certainty. To this end micrometeorological and hydrological measurements were made over one growing season using automatic weather stations and weighing micro-lysimeters (18,716 cm^3) at several locations within a multiple land-use agricultural catchment in Southern Ontario. This paper compares modelled equilibrium evapotranspiration (PET_{EQ}) and measured actual evapotranspiration (AET) values obtained from the lysimeters in riparian zone in a multiple land-use agricultural watershed in Southern Ontario. Two sites were chosen in two different riparian areas of the watershed, representing the range in surface conditions dominant throughout the basin.

The results yield a mean daily AET of 3.43 mm for the catchment's riparian zone, with virtually no differences between the upper and lower portions of the basin, which lost 3.36 and 3.49 mm d^{-1}, respectively. Using these daily AET values along with coincident estimates of PET_{EQ} produced Priestley–Taylor coefficients (α) of 1.10 and 1.18 at the upper and lower sites, respectively.

1. Introduction

Evapotranspiration (ET) is extremely important in the water balance of agricultural watersheds as it can affect moisture availability for crops as well as soil water storage potential and therefore overall catchment runoff. Previous research on ET from agricultural areas has focused on croplands and little attention has been given to ET from riparian areas despite their importance (Bowie and Kam, 1968; Federer, 1973; Unland et al., 1998). Riparian zones found along river and stream corridors in agricultural basins have the function of regulating stream water quantity and quality (MacNish et al., 2000). From a basic water balance perspective, the quantification of the atmospheric exchange of water through ET from the riparian zone is essential to the basin-wide water balance and the groundwater recharge/depletion process, critical to the upslope moisture status available to crops (Maddock et al., 1998). In basins such as the study basin, groundwater

* *Corresponding author*. Tel.: +1 519 884 0710; fax: +1 519 725 1342.
E-mail address: rpetrone@wlu.ca (R.M. Petrone).
0378-3774/$ – see front matter © 2006 Elsevier B.V. All rights reserved.
doi:10.1016/j.agwat.2006.05.018

contributions to the basin hydrology can be significant (Mengis et al., 1999; Macrae et al., in press-a), and as such groundwater models are often an integral part of the management of water resources for agricultural and human consumption. Such models require accurate estimates of basin boundary conditions, the most important of which is ET (Goodrich et al., 2000). Riparian zone contributions to basin ET will be a function of the width (areal extent) of the zone, which in systems with large floodplains can represent a significant portion of seasonal precipitation and may be critical from a water use planning perspective. Thus, an estimate of ET loss from the riparian zone should also be considered in irrigation planning.

From a water quality perspective, ET plays an important role by drying out surface soils between precipitation events. The importance of wetting-drying cycles on soil nutrient availability has been demonstrated in numerous studies (e.g. Pote et al., 1999; Macrae et al., in press-b). In addition to nutrient cycling, nutrient retention and/or transport in wetlands of any type depends on the hydrologic balance, which is dominated by ET (Lafleur, 1990; Petrone et al., 2004a,b; Andersen et al., 2005). Therefore, to quantify the nutrient exchange within an agricultural basin the water balance of the riparian zones (in addition to the cropped fields) must be quantified—especially the factors controlling ET.

Regardless of these two crucial aspects of riparian zones in the hydrological functioning of an agricultural basin, ET is rarely known with great certainty (Goodrich et al., 2000), especially its seasonal rates and degree of spatial variability (Goodrich et al., 2000; Maddock et al., 1998; Unland et al., 1998; Chehbouni et al., 2000). Much of this lack of information is due to none of the typical micrometeorological approaches for measuring ET being appropriate for the narrow riparian corridor (Chehbouni et al., 2000). One approach to overcoming this challenge is to base AET measurements on some estimation of evaporative power or demand (potential or potential equilibrium evapotranspiration) or weighing microlysimeters (Hillel, 1998).

As such, many studies are concerned with potential evapotranspiration (PET), which is the maximum rate of ET from a vegetated catchment under the condition of unlimited moisture supply and without advection or heat storage effects (Thomas, 2000; Jacobs et al., 2002). Estimating evaporation by this approach has the advantage of being relatively simple and direct, and makes use of easily measured atmospheric, soil and vegetation variables (Granger and Gray, 1989). Furthermore, PET is a commonly used approach to provide efficient, cost effective estimates of evaporative losses for agricultural and engineering applications (Brutsaert, 1988).

While the FAO-Penman–Monteith (P–M) model is currently the accepted standard for predicting crop reference ET (Kashyap and Panda, 2001), this paper uses the Priestley–Taylor (P–T) method for estimating PET and AET values obtained from soil weighing lysimeters to determine the range in coefficients representative of riparian zone surfaces in a temperate agricultural watershed. The primary reason for not using the FAO-P–M model in this study is that this model is intended to produce a reference ET. This concept is based on a hypothetical crop with an associated height of 0.12 m, r_s of 70 s m^{-1} and an α of 0.23, which is not very representative of a riparian system (Andersen et al., 2005). The Priestley–Taylor model has also been shown to outperform P–M in wetland/grass sites as it simulates ET well, is computationally more efficient, and does not require the collection of wind speed and relative humidity data (Summer, 1996). P–M and P–T have also compared well during seasons in agricultural sites (Eitzinger et al., 2002). Further, in the hydrological and climatological communities, alpha (α, obtained via P–T) is often used in process-based studies as a valuable tool to gain some insight into the degree of moisture stress, and/or the degree of water conservation by vegetation (Petrone et al., 2004a,b; Eichinger et al., 1996; Davies and Allen, 1973). Having a reliable α value is useful as it may then be used in more detailed PET models like the P–M to obtain estimates of actual ET. That is, as outlined in classic papers such as Davies and Allen (1973) alpha can be directly related to P–M PET to not only obtain actual ET but also model r_s.

Thus, the objectives of this study are three fold: (1) to provide estimates of summer mean daily ET rates from narrow, grassed riparian areas using micro-lysimeters and the P–T method; (2) to provide good estimates of α for riparian vegetation throughout an agricultural basin on daily scales; and (3) to examine spatiotemporal variability in riparian zone ET and α. An improved understanding of spatial and temporal patterns in riparian ET is valuable to the management community as it provides insight on the vulnerability of riparian soils to drying, as this in turn affects the quality and quantity of agricultural runoff passing through these zones into surface waters. Furthermore, the α values provided by this study can be used to improve models of overall catchment water balances.

2. Study site

This study was conducted in the Strawberry Creek watershed, a small multiple land-use agricultural watershed located in Southern Ontario. Strawberry Creek is a perennial, first order stream located in Maryhill, Ontario, approximately 20 km northeast of Waterloo, Ontario (42°33′N, 80°23′W) (Fig. 1). The creek is approximately 2 km in length and drains a watershed of about 3 km^2 (Harris, 1999). Strawberry Creek then flows eastward into Hopewell Creek, which subsequently drains into the Grand River through to Lake Erie.

Strawberry Creek is located in a humid continental climate region (Brown et al., 1980). There are approximately 3200 growing degree-days and 130 frost-free days with a mean annual temperature of 6.7 °C (Brown et al., 1980). The mean annual precipitation is 858 mm, with approximately 16% of precipitation falling as snow. The mean annual PET for this region is approximately 590 mm, and mean annual AET is approximately 558 mm (Brown et al., 1980).

The Strawberry Creek watershed contains a variety of vegetation types. The agricultural fields contain soybean (*Heterodera glycines*), and corn (*Zea mays*). In the lower portion of the watershed approximately 60% of the arable land has been in fallow for 9 years, the remaining fraction is under corn and strawberry cultivation. In the upper portion of the watershed corn and soybean are grown. At the headwaters of Strawberry Creek is a deciduous swamp and there are two smaller woodlots within the watershed. Streamflow is

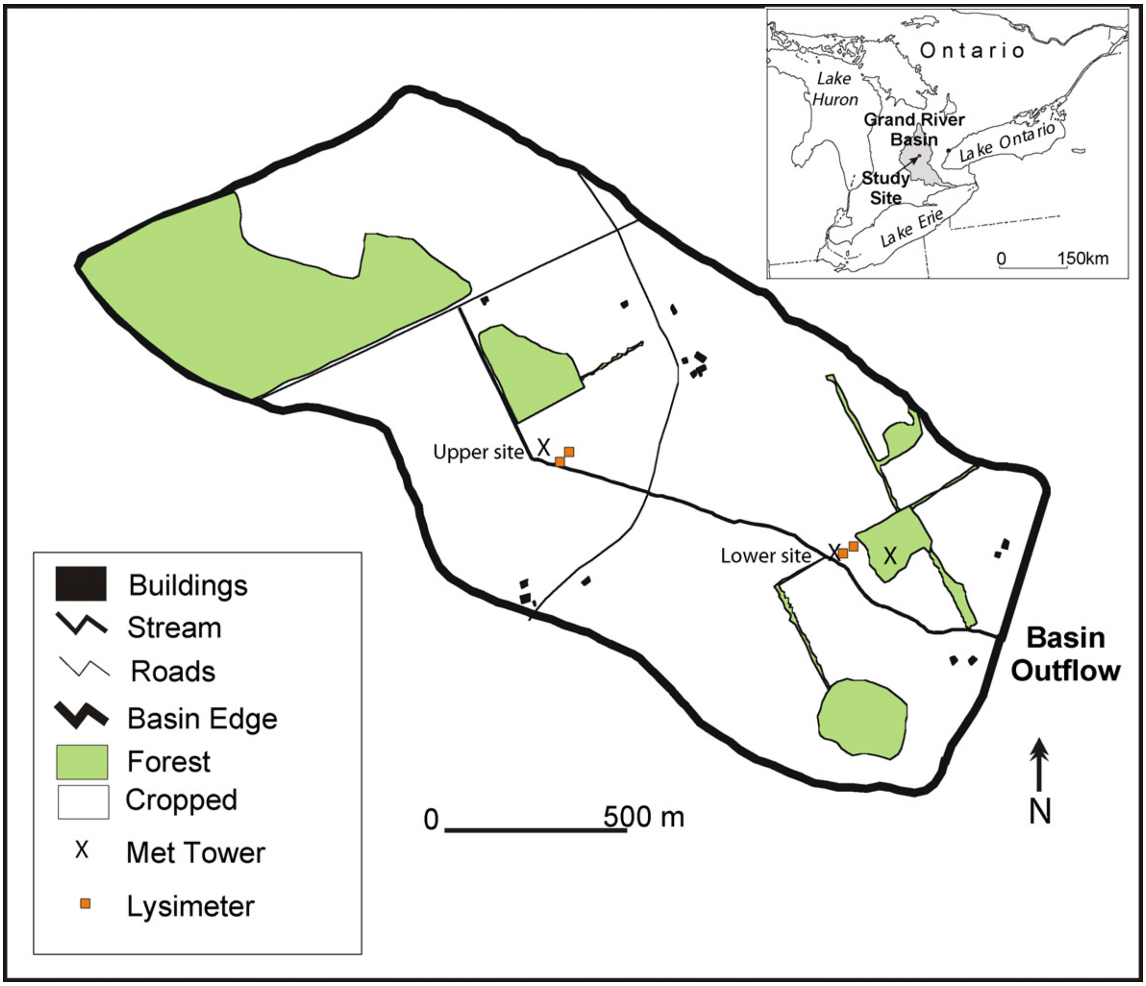

Fig. 1 – Schematic representation of the Strawberry Creek basin, showing the location of instrumentation and land-use types. Inset: location of the Strawberry Creek study site, relative to Southern Ontario.

sustained by two deciduous swamps, nine drainage tiles, and a perennial groundwater source. The riparian areas that line Strawberry Creek are 10–15 m wide and of low gradient. These riparian zones are dominated by tall grasses (*Graminae* spp.) but also contain Canada Goldenrod (*Solidago canadensis*), Yarrow (*Achillea millefoleum*), Thistles (*Cirsium* spp.), Common Milkweed (*Asclepias syriaca*), Queen Anne's Lace (*Daucus carota*), Common Burdock (*Arctum minus*), and a few trees (primarily *Salix* spp.).

The riparian zone at Strawberry Creek is fairly homogeneous although there is some micro-topographic variability. Two measurement sites were established in the watershed that are relatively similar and are representative of surface conditions observed within the riparian zone (Fig. 1). The first is located in the northern end of the basin, and will be herein referred to as the ''upper site''. Here the vegetation consists of riparian flora (mainly grasses but no trees) and the soils at this site are characterized as a sandy loam with the top 30 cm of soil having an average bulk density of 0.85 ± 0.16 g/cm^3, with a bulk density of 1.14 ± 0.16 g/cm^3 at the 30–40 cm depth. The upper layers of the soil (0–30 cm) had an average organic content of approximately 24%, while the bottom layer had an organic content of only 8%. The upper site is also closest to the headwaters of Strawberry Creek where stagnant water frequently pools on the adjacent cornfield, which would help to slow decomposition, increasing the carbon content and porosity of the upper soil (Brady and Weil, 1999).

The second study site, referred to herein as the ''lower site'', is located in the southern end of the basin and contains riparian vegetation and the soil at this site is also characterized as a sandy loam, with an average bulk density of approximately 1.06 ± 0.09 g/cm^3 in the upper 30 cm, and a porosity of 53%. The soils at the lower site were composed of 8% organic material over most of the profile.

3. Methodology

Lysimeters have been used extensively to measure the net movement of water across the soil-atmosphere boundary (Tanner, 1967; Seyfried et al., 2001). A lysimeter is an artificially enclosed volume of soil that can be placed in the field and filled with representative soil and/or vegetation (Brutsaert, 1988). For this study weighing micro-lysimeters were constructed from a 40 cm $\times$ 26 cm $\times$ 18 cm rectangular enclosure (18,716 cm^3) nested inside a second plastic container. Material from the surface (including soil and vegetation) was extracted (by block cutting) and placed in the first enclosure. In order to maintain the same mechanical properties of the soil, it was placed in the bucket as an undisturbed block (monolith), as suggested by Brutsaert (1988). The enclosure containing the soil/plant monolith had holes drilled (at a density of nine 3 mm holes per 4 cm^2) into the bottom to facilitate the drainage (percolation) of water. The second container was not perforated and contained the first container to collect its drainage, the volume of which was measured each time the lysimeter was weighed. Each lysimeter was placed in the hole from which the monolith was extracted.

At the upper and lower sites triplicate lysimeters of bare soil and vegetation were installed for a total of 12 lysimeters within the catchment's riparian zone (Fig. 1). The first location was at the northern end (upper site) of the basin located in a riparian area adjacent to a cornfield. The second location was also in a riparian area (adjacent to a soybean field) in the lower portion (lower site) of the basin.

These smaller weighing lysimeters have been used successfully in other studies (cf. Van Seters and Price, 2001; Petrone et al., 2004a,b) to obtain estimates of ET that were comparable with other micrometeorological approaches. Further, they offer the most significant advantage over larger more elaborate lysimeters in that they are more cost effective and easier to install and maintain more spatial replication of measurements. Their draw back is that they can be a challenge to consistently mirror the natural conditions observed outside of the lysimeters. To overcome this challenge more frequent measurements are required of not only the weight of the lysimeter but also the volume of percolated water and water input to the system. Water input to the lysimeter includes precipitation, surface runoff and any irrigation required to bring the lysimeter soils up to similar soil moisture values as the soil that surrounds it. Thus, regular TDR measurements were taken inside and outside of the lysimeter each time a measurement was taken (Van Seters and Price, 2001; Petrone et al., 2004a,b).

Furthermore, there may be concern that a lysimeter with a depth of 26 cm may sever root systems leading to a decrease in transpiration within it. Observations of the lysimeters while in use suggested little mortality relative to the vegetation surrounding the lysimeter. In addition, approximately 75% of all the roots of most of the vegetation identified in these riparian sites are located within the top 30 cm (Wynn et al., 2003, 2004).

Each lysimeter was weighed with a precision Chatillon® spring balance (Ametak Inc.). The weight of each lysimeter enclosure was recorded and the water that had drained (if any) into the bottom enclosure was also measured by emptying the contents into a 1 l graduated cylinder. Soil moisture was also recorded in each lysimeter, and in the immediate area surrounding the lysimeter using a portable time domain reflectometry (TDR) instrument (Hydrosense Soil Moisture Probe, Campbell Scientific Ltd.). This allowed the soil moisture to be monitored to ensure that the soil in the lysimeters was comparable to the surrounding soil. AET from lysimeters is quantified via:

$$\mathrm{AET} = \frac{P - (V_L + V_R + \Delta V_S)}{A} \tag{1}$$

where P is the precipitation (mm), V_L the volume of drainage loss (mm^3), V_R the volume of net surface water exchange (mm^3), ΔV_S the change in the water storage in the lysimeters (mm^3) and A is the area of the lysimeter (mm^2) (Tanner, 1967). ΔV_S was measured by collecting the water that drained through the contents of the lysimeter and differences in weight of the lysimeter. Lysimeter AET obtained in this manner is then compared with modelled PET values for the same sites.

Automated meteorological stations were also installed at the upper and lower sites to continuously monitor net radiation, air temperature and relative humidity, wind speed and direction, ground heat flux, and ground temperatures. Only the lower station was equipped with a tipping bucket rain gauge, which was used to record basin precipitation to be used in the lysimeter calculations. Soil moisture TDR probes were also buried 10 and 50 cm below the surface at both sites to assess moisture conditions.

A vegetation count and survey was conducted in the immediate vicinity of each lysimeter and meteorological station. The area of the vegetation survey was determined by the height of the net radiometer, and the ratio of 10:1 m, where a circular area (or tower footprint) with a radius of 10 m is observed by a radiometer at a height of 1 m (Oke, 1987). The height of the radiometers were 1.5 and 2 m at the lower and upper sites, respectively. Therefore, a radius of 15 and 20 m was used at the lower and upper sites, respectively. The vegetation area within these footprints was recorded along with the coverage of bare soil and then used to areally weight the lysimeter values.

The Priestley–Taylor equation is classified as a radiation-based approach to estimating ET, using net radiation and air temperature to evaluate equilibrium evaporation, which assumes that an air mass moving over a homogeneous, well-watered surface would become saturated (Dingman, 1994; Priestley and Taylor, 1972). Under these ideal conditions ET would eventually reach a rate of equilibrium (equilibrium potential evapotranspiration, $\mathrm{PET_{eq}}$) (Priestley and Taylor, 1972). Radiation is a very effective parameter to use in measuring equilibrium evaporation or PET. In a review of 30 studies it was commonly found that in vegetated areas with very small, or no, water deficits approximately 95% of the annual evaporative demand was supplied by radiation (Stagnitti et al., 1989). The Priestley–Taylor model obtains $\mathrm{PET_{eq}}$ via:

$$\mathrm{PET_{eq}} = \frac{\Delta}{\Delta - \gamma}(Q^* - Q_G) \tag{2}$$

where Δ is the slope of the saturated vapour pressure curve (°C/kPa), γ the psychometric constant (Pa K^{-1}), Q^* the net radiation (W m^{-2}) and Q_G is the soil heat transfer (W m^{-2}). $\mathrm{PET_{eq}}$ is related to AET via the Priestley–Taylor coefficient (α). The term α is generally solved using:

$$\alpha = \frac{\mathrm{AET}}{\mathrm{PET_{eq}}} \tag{3}$$

where AET is the total measured evaporation and $\mathrm{PET_{eq}}$ is the total equilibrium evaporation (Wilson and Baldocchi, 2000). However, equilibrium rarely occurs, as there is almost always horizontal advection and deviations from a wet surface

(Wilson and Baldocchi, 2000). In many studies 1.26 is the value used for α, which is based on conditions of minimum advection and no edge effects (Dingman, 1994; Jacobs et al., 2002). Under such conditions potential evapotranspiration is equal to the equilibrium evapotranspiration (Davies and Allen, 1973; Eichinger et al., 1996). Alternative general estimates for α have also been made with respect to certain surfaces. For example, Morton (1983) suggested that an α value of 1.32 should be used for vegetated surfaces due to surface roughness, and in arid regions α should be increased to 1.74. However, a daily average value of 1.26 has been found to be appropriate for most humid climates (Priestley and Taylor, 1972; Rouse and Stewart, 1972).

4. Results and discussion

4.1. Climatology and basin hydrology

This study spanned the warm, dry summer months. The temperature of the basin is presented as an arithmetic mean of temperature data from three meteorological stations placed throughout the basin (Fig. 1). The mean (and standard deviation) temperature of the basin over the study period was $18.9 \pm 2.6\,^{\circ}C$, which is comparable to the 30-year normal of $18.6\,^{\circ}C$ over the same 3-month period (Environment Canada, 2004). The temperature over the study period ranged from a low of $12.0\,^{\circ}C$ on 4 June, to a high of almost $24.0\,^{\circ}C$ on 4 July, after which temperatures generally decreased over the course of the season (Fig. 2a). However, during this period of decline a second maximum of $22.8\,^{\circ}C$ was recorded for a brief period on 13 August.

As observed with temperature, the two sites followed a similar seasonal pattern for relative humidity (RH) with mean values and standard deviations of $82.4 \pm 19.3\%$ and $77.0 \pm 17.9\%$ at the upper and lower sites, respectively (Fig. 2e). Daily averages for the lower site ranged from a minimum of 41.8% to a maximum of 100%. For the upper site, the daily values ranged from 51.9% to 94.9%.

A total of 178 mm of precipitation was recorded over the study period, with the largest single day total of 16.5 mm occurring on 15 July (Fig. 2f). This is considerably lower than the 30-year normal of 259.4 mm over the same time period (Environment Canada, 2004). Precipitation in the basin was classified into three wet periods separated by distinct drier periods noted by a lack of rainfall (Fig. 2f). The three wet periods spanned 2 June to 18 June, 5 July to 21 July and 2 August to 16 August, respectively.

Basin discharge at the stream outflow was largely fed by slow ground water flow over the study period. Tiles did not flow over the duration of the study and no surface inputs of runoff were observed. Basin discharge over the study period was low compared to what is generally observed over the remainder of the year, but was comparable to what has been observed in the summer months in the past (Macrae et al., in press-a). Increases in discharge for the basin followed the large precipitation events at the start and end of the study period. In general, throughout much of the middle of the study period the discharge was low and responses to periodic precipitation events were minimal. However, discharge from the middle wet period did not respond in the same manner as the other two rain events (Fig. 2d). This lack of a response in discharge can be explained by dry soil moisture conditions in the basin during that period, where much of the precipitation went into storage or AET. Soil moisture data indicate that both sites had a significant potential for moisture to be stored in the soil during the middle of the season (Fig. 2b and c). Fig. 2b and c shows that the upper soil layers were much drier than the deeper layers during this middle period. Thus, during this time the soils had dried enough that the middle wet period produced no response in soil moisture in the lower layer, only the upper layer was affected as soil moisture storage was replenished. During this middle wet period the rain infiltrated rapidly to raise the water table, rather than entering the stream as discharge. These results demonstrate the importance of cumulative daily ET at creating dry antecedent hydrologic conditions in soils, which allow riparian soils to retain runoff instead of allowing it to pass into the stream.

Towards the end of the study period the soil moisture in the upper layers of the soil increased at both the upper and lower sites (Fig. 2b and c). Over a 4-day period (2–6 August) the soil moisture increased from 0.22 to 0.38 and 0.21 to 0.35 $cm^3\,cm^{-3}$ at the lower and upper sites, respectively. This was followed by a decrease during a reprieve in the precipitation, and then an increase to the highest moisture level of the study period in response to the second largest rain event of the season (12 August). The lower site reached maximum values of 0.40 and 0.43 $cm^3\,cm^{-3}$ for the deeper and shallower layer, respectively, while the upper site reached 0.39 and 0.44 $cm^3\,cm^{-3}$ for the deeper and shallower layer, respectively (Fig. 2b and c).

The data show that the amount of moisture that could be stored in the soil was highest in the middle of the season at both sites. Immediately prior to this period, there was a lack of rain allowing the soil to dry out. During the next wet period, the precipitation was able to infiltrate the soil fairly rapidly and remain there, rather then being discharged throughout the basin as lateral groundwater flow (Fig. 2d). During this second wet period the soil moisture graphs indicate that there was more potential for moisture storage at the upper site than at the lower site, which can be attributed to the soil properties of the site (high porosity and a high organic content). This also suggests that more water was available at the upper site for AET. However, the α values presented below do not support this.

4.2. Energy balance

Fig. 3a and b shows similar net radiation (Q^*) and ground heat flux (Q_G) values for the two sites. The mean Q_G was 0.09 and 0.10 $W\,m^{-2}$ at the lower and upper sites, respectively, and ranged from approximately -0.2 to 0.55 $W\,m^{-2}$ at both sites over the study period. Mean Q^* values were 106.3 and 100.2 $W\,m^{-2}$ at the lower and upper sites, respectively. Both the Q^* and Q_G values at the two sites showed little variability over the study season, but exhibited strong daily fluctuations. The lack of a clear seasonal trend is the result of the study period not spanning the shoulder seasons (spring and autumn).

Wind speed varied the most between the two sites, and was much higher at the upper site than at the lower site (Fig. 3d). The average wind speed and standard deviation at the lower

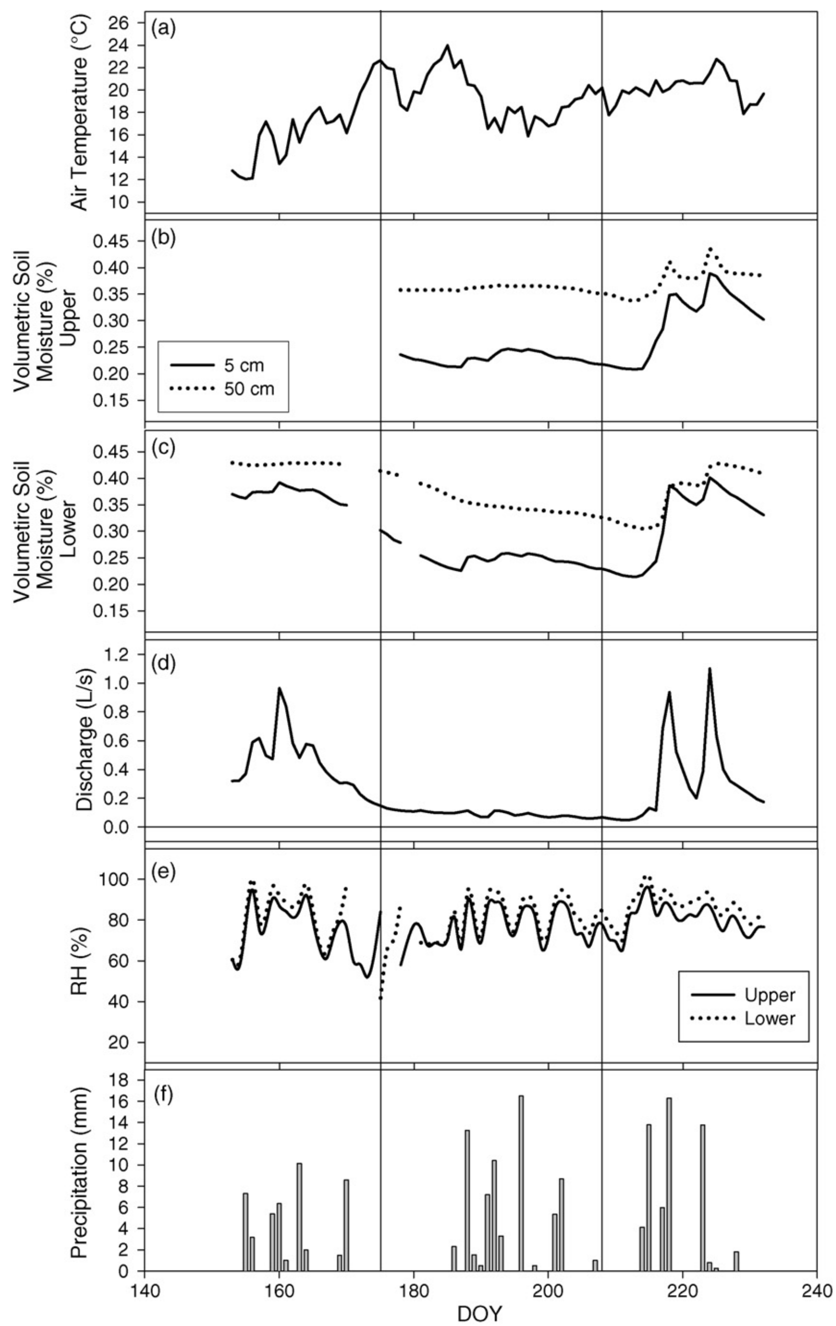

Fig. 2 – Basin hydrology and climatology for Strawberry Creek, 3 June to 22 August 2003, Maryhill, Ontario. (a) Mean air temperature for the basin, (b) volumetric soil moisture (%) at the upper site, (c) volumetric soil moisture (%) at the lower site, (d) discharge of the basin (mm), (e) relative humidity (%) and (f) precipitation (mm). Vertical lines delineate the three wet periods.

site was 1.5 ± 1.1 m s^{-1}, and the upper site was 1.7 ± 1.4 m s^{-1} (a difference of 13%). However, the maximum daily wind speed was 3.2 and 5.7 m s^{-1} for the lower and upper sites, respectively. Furthermore, the upper site wind gusts were observed to exceed 9 m s^{-1}, while at the lower site the highest wind gust observed was 6.6 m s^{-1}.

4.3. *Evapotranspiration*

The total summer PET (and standard error) for the lower site was 201.1 ± 3.98 mm, and for the upper site was 205.5 ± 3.69 mm. Average daily PET (and standard error) rates of 2.9 ± 0.1 and 3.1 ± 0.1 mm d^{-1} were measured at the lower and upper sites, respectively. The AET losses at the two sites, obtained using the lysimeters, were also very similar. The lower site had a total AET (with standard error) of 237.1 ± 3.9 mm, and the upper site had a total of 223.8 ± 5.4 mm. These seasonal AET rates translated to average daily AET rates of 3.49 and 3.36 mm d^{-1} for the lower and upper sites, respectively. At both sites AET from the vegetation contributed to the bulk of the AET (Table 1). AET rates obtained in this study are comparable with other studies

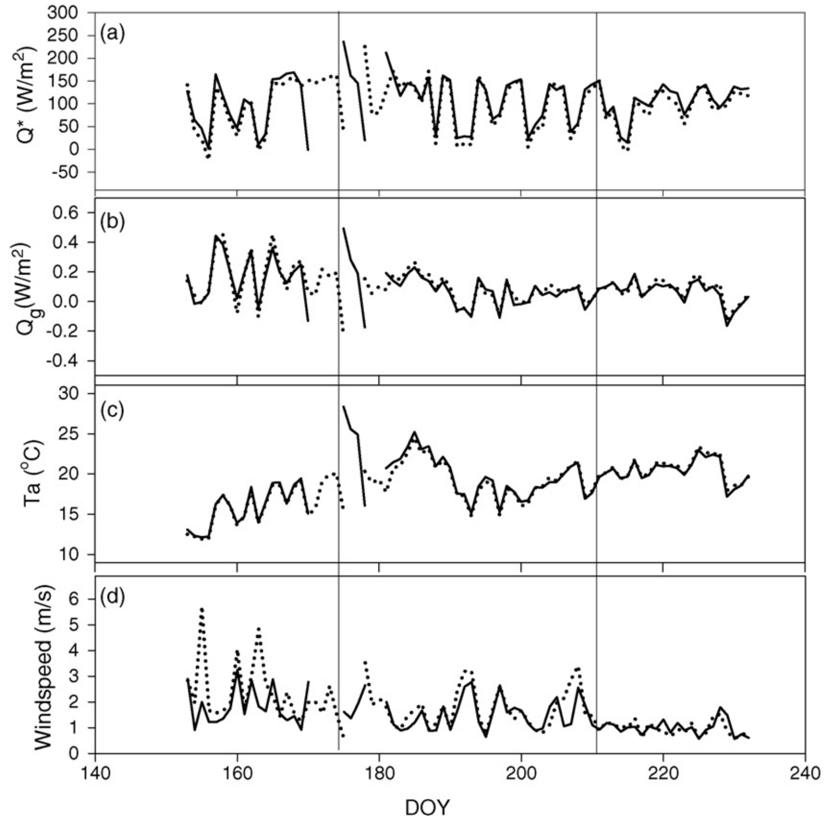

Fig. 3 – Energy balance and micrometeorological components for the study period 3 June to 22 August 2003, Strawberry Creek, Maryhill, Ontario. (a) Net radiation (W m^{-2}), (b) ground heat flux (W m^{-2}), (c) temperature (°C), and (d) wind speed (m s^{-1}). Data from the lower site are represented by a solid line and the upper site by a dotted line. The vertical lines delineate the wet periods.

on similar agricultural grass and riparian areas. The daily AET rates of 3.49 and 3.36 mm d^{-1} for both sites are well within the range of 0.5–4 mm d^{-1} reported in a similar terrain by Unland et al. (1998) although they are at the higher end of the range.

The α coefficient was determined for both sites using the lysimeter data along with energy balance measurements. The α values used are a compilation of all the vegetation in the area and the soil, weighted accordingly to their respective land cover within the footprint of the respective net radiometers.

Table 1 – Areally weighted seasonal mean daily actual evapotranspiration (AET) from the upper and lower sites, Strawberry Creek watershed, Maryhill, Ontario for the 29 measurement days, spanning DOY 153–232, 2003

	Upper site (mm d^{-1})	Lower site (mm d^{-1})
Soil cover	0.14 (0.02)	0.19 (0.04)
Vegetation cover	3.22 (0.35)	3.30 (0.71)
Total weighted	3.36 (0.73)	3.49 (0.71)

Means are based on date collected from three bare soil and three vegetated lysimeters at each of the sites, areally weighted according to the respective distributions within the footprint of the two radiation sensors. Standard errors are shown in parentheses.

While few have focused primarily on riparian areas the range reported here (1.10–1.18) is also well within the range of 0.06–2.10 presented for various agricultural grass-dominated sites (Ritchie, 1972; Wallace and Holwill, 1997; Katal and Parlange, 1992; Hares and Novak, 1992). Over the study period, α ranged from 1.06 to 1.25 at the lower site, and 1.05–1.15 at the upper site. The seasonal average α was essentially the same at both sites, with values of 1.18 and 1.10 at the lower and upper sites, respectively. From a modelling perspective these results illustrate that α estimates of 1.26 are quite reasonable for the mid-summer period in a temperate riparian area such as this study site. These results also suggest that in a first order agricultural catchment riparian zone ET modelling is not spatially variable, potentially significantly simplifying the spatial aggregation of models such as P–T or FAO-P–M. Thus, in management and modelling applications it is safe to consider riparian zones as one homogeneous landscape unit in terms of generating ET estimates.

Although PET, AET and alpha did not differ spatially within riparian areas, some temporal variability was observed (Table 2). Using the clear wet periods illustrated in Fig. 2f, three seasonal periods (DOY 153–175, 176–208, 209–232) were identified to examine any seasonal trends. Both sites showed maximum AET and PET amounts in the middle period (DOY 176–208) (Table 2). The maximum α value at both sites was observed in the first period, but the upper site α remained

Table 2 – Period daily averages in actual evapotranspiration (AET) and potential equilibrium evaporation (PET), and the resulting Priestley–Taylor coefficient (α) for the upper and lower sites, Strawberry Creek watershed, Maryhill, Ontario, DOY 153–232, 2003

DOY	Number of samples	Upper site			Lower site		
		AET	PET	α	AET	PET	α
153–175	8	3.25	2.77	1.17	3.05	2.56	1.19
176–208	11	4.08	3.26	1.25	4.07	3.01	1.35
209–232	10	2.74	3.16	0.87	3.36	3.29	1.02

Periods are defined by the wet periods delineated in Fig. 2. AET and PET are the areally weighted totals for both sites as described above.

larger than that at the lower site in the third period (DOY 209–232) (Table 2). The fact, that the α values differed little between the two sites should not come as a surprise as a dominant controlling factor on α is moisture availability (Dingman, 1994) and precipitation and soil moisture were similar between the lower and upper sites.

A total of 178 mm of precipitation fell over the study period whereas slightly over 100 mm of ET was observed from riparian areas (i.e. nearly 60%). To put these numbers into perspective, less than 2 mm of runoff was observed at the basin outflow over the entire study period. Therefore, the water balance of the basin was dominated by atmospheric exchange during the study period. While these dry periods do not contribute significantly to annual basin hydrochemical export as most nutrient loading occurs during storm events (Macrae et al., in press-a), such periods are critical to nutrient loading patterns because the cumulative daily ET that occurs during these periods dries out riparian soils affecting both their water storage and nutrient retention potential. Thus, ET patterns during the dry summer months have a direct effect on the ability of riparian zones to respond to runoff and its associated nutrient transport during late summer and autumn storms.

5. Conclusions

Estimates of summer PET and AET for riparian zones within a first-order agricultural catchment were 2.9–3.1 and 3.36–3.49 mm d^{-1}, respectively. Nearly 60% of summer precipitation was returned to the atmosphere via ET in riparian areas and thus atmospheric exchange dominated the water balance of these zones during the study period. In basins with larger, more extensive riparian zones and floodplains, these large AET rates may account for a significant portion of the water available to the entire basin, which has implications for irrigation planning.

Estimates of the P–T coefficient (α) were 1.10–1.18 over the season supporting the use of a value of 1.26 that has been suggested by others for use in modelling scenarios. Furthermore, no spatial variability in ET was observed between sites within riparian areas suggesting that these areas may be treated as one homogenous unit with respect to ET.

In small basins such as the Strawberry Creek watershed, the cumulative ET during dry periods plays a critical role in drying riparian soils between storms. These dry antecedent hydrologic conditions allow storm runoff to go into storage rather than passing through riparian areas into the adjacent stream, and thus affect both stream water quality and quantity. Therefore, it is critical to understand this loss of water in a riparian zone to ensure the hydrologic conditions necessary to maintain the critical biogeochemical functioning of the riparian zone.

Riparian zones are often ignored in studies of ET in agricultural watersheds. However, as this study illustrates, less intensive research methods could be used such as micro-lysimeters and simple energy balance measurements, which can be carried out using automated instruments, once relationships between PET and AET are quantified for the various representative land-use types comprising the system. Such relationships can then be used as more meaningful functional relationships to parameterize more detailed models like the FAO-P–M.

Acknowledgements

Funding for this research was provided by the Natural Science and Engineering Research Council of Canada, the Canadian Foundation for Innovation, and a Wilfrid Laurier University Short-term Research Grant. The field assistance of Patrick Chahil and J.R. Van Haarlem is gratefully acknowledged. The authors wish to thank the anonymous reviewers and the Joint Editor in Chief for their insightful comments.

REFERENCES

Andersen, H.E., Hansen, S., Jensen, H.E., 2005. Evapotranspiration from a riparian fen wetland. Nord. Hydrol. 36 (2), 121–135.

Bowie, J.E., Kam, W., 1968. The use of water by riparian vegetation, Cottonwood Wash, Arizona. US Geological Survey Water Supply Paper No. 1858.

Brady, N.C., Weil, R.R., 1999. The Nature and Properties of Soils. Prentice Hall, USA.

Brown, D.D., McKay, G.A., Chapman, L.J., 1980. The Climate of Southern Ontario. Environment Canada, Toronto.

Brutsaert, W., 1988. Evaporation into the Atmosphere: Theory, History and Applications. D. Reidel Publishing Company, Dordrecht.

Chehbouni, A., Goodrich, D.C., Moran, M.S., Watts, C.J., Kerr, Y.H., Dedieu, G., Kepner, W.G., Shuttleworth, W.J., Sorooshian, S., 2000. A preliminary synthesis of major scientific results during the SALSA program. Agric. For. Meteorol. 105, 311–323.

Davies, J.A., Allen, C.D., 1973. Equilibrium, potential and actual evaporation from cropped surfaces in Southern Ontario. J. Appl. Meteorol. 12 (4), 649–657.

Dingman, S.L., 1994. Physical Hydrology. Prentice Hall Inc., New Jersey.
Eichinger, W.E., Parlange, M.B., Stricker, H., 1996. On the concept of equilibrium evaporation and the value of the Priestley–Taylor coefficient. Water Resour. Res. 32 (1), 161–164.
Eitzinger, J., Marinkovic, D., Hösch, J., 2002. Sensitivity of different evapotranspiration calculation methods in different crop–weather models. In: Proceedings of the International Modelling and Software Society Meeting (IEMSS 2002), vol. 2, Lugano, Switzerland, pp. 395–400.
Environment Canada, 2004. Canadian Climate Normals 1971–2000. Environment Canada National Climate Archive, Toronto.
Federer, C.A., 1973. Forest transpiration greatly speeds streamflow recession. Water Resour. Res. 9 (6), 1599–1604.
Goodrich, D.C., Scott, R., Qi, J., Goff, B., Unkrich, C.L., Moran, M.S., Williams, D., Schaeffer, S., Snyder, K., MacNish, R., Maddock, T., Pool, D., Chehbouni, A., Cooper, D.I., Eichinger, W.E., Shuttleworth, W.J., Kerr, Y., Marsett, R., Ni, W., 2000. Seasonal estimates of riparian evapotranspiration using remote and in situ measurements. Agric. For. Meteorol. 105, 281–309.
Granger, R.J., Gray, D.M., 1989. Assessment of Morton's Model for Estimating Areal Evapotranspiration. Environment Canada, Saskatoon.
Hares, M.A., Novak, M.D., 1992. Simulation of surface energy balance and soil temperature under strip tillage. II. Field test. Soil Sci. Soc. Am. J. 56, 22–29.
Harris, M., 1999. Nitrate attenuation in a narrow non-forested riparian buffer zone in an agricultural watershed in Southern Ontario. MES Thesis. Wilfrid Laurier University, Waterloo.
Hillel, D., 1998. Environmental Soil Physics. Academic Press, USA.
Jacobs, J.M., Mergelsberg, S.L., Lopera, A.F., Myers, D.A., 2002. Evapotranspiration from a wet prairie wetland under drought conditions: Paynes Prairie Preserve, Florida, USA. Wetlands 22 (2), 374–385.
Kashyap, P.S., Panda, R.K., 2001. Evaluation of evapotranspiration estimation methods and development of crop-coefficients for potato crop in a sub-humid region. Agric. Water Manage. 50, 9–25.
Katal, G.G., Parlange, M.B., 1992. Estimation of bare soil evaporation using skin temperature measurements. J. Hydrol. 132, 91–106.
Lafleur, P.M., 1990. Evapotranspiration from sedge-dominated wetland surfaces. Aquat. Bot. 37, 341–353.
MacNish, R.D., Unkrich, C.L., Smythe, E., Goodrich, D.C., Maddock III, T., 2000. Comparison of riparian evapotranspiration estimates based on a water balance approach and sap flow measurements. Agric. For. Meteorol. 271–279.
Macrae, M.L., English, M.C., Schiff, S.L., Stone, M.A., in press-a. Temporal patterns of nutrient export from a first-order agricultural basin in Southern Ontario. Hydrol. Process.
Macrae, M.L., Creed, I.F., Macdonald, S.E., Devito, K.J., in press-b. Relation of soil nitrogen distribution and surface and ground water nitrogen concentrations in harvested and unharvested portions of an aspen-dominated catchment in the Boreal Plain. Can. J. For. Res.
Maddock III, T., MacNish, R.D., Goodrich, D.C., Williams, D.G., Shuttleworth, W.J., Goff, B.A., Scott, R.L., Moran, M.S., Cooper, D.I., Hipps, L.E., Chehbouni, A.G., 1998. An overview of atmospheric and surface water coupling to regional groundwater models in semi-arid basins. In: Proceedings of the American Meteorological Society of Special Symposium on Hydrology, Phoenix, AZ, pp. 38–42.
Mengis, M., Schiff, S.L., Harris, M., English, M.C., Aravena, R., Elgood, R.J., MacLean, A., 1999. Multiple geochemical and isotopic approaches for assessing ground water NO_3 elimination in a riparian zone. Ground Water 37, 448–457.
Morton, F.I., 1983. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology. J. Hydrol. 66, 1–76.
Oke, T.R., 1987. Boundary Layer Climates, 2nd ed. Routledge, Cambridge.
Petrone, R.M., Price, J.S., Carey, S.K., Waddington, J.M., 2004a. Statistical characterization of the spatial variability of soil moisture in a cutover peatland. Hydrol. Process. 18 (1), 41–52.
Petrone, R.M., Price, J.S., Waddington, J.M., von Waldow, H., 2004b. Surface moisture and energy exchange from a restored peatland, Québec, Canada. J. Hydrol. 295, 198–210.
Pote, D.H., Daniel, T.C., Nichols, D.J., Moore Jr., P.A., Miller, D.M., Edwards, D.R., 1999. Seasonal and soil-drying effects on runoff phosphorus relationships to soil phosphorus soil. Sci. Soc. Am. J. 63, 1006–1012.
Priestley, C.H.B., Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large scale parameters. Mon. Weather Rev. 100, 81–92.
Ritchie, J.T., 1972. Model for predicting evaporation from a crop with incomplete cover. Water Resour. Res. 8, 1204–1213.
Rouse, W.R., Stewart, R.B., 1972. A simple model for determining evaporation from high-latitude upland sites. J. Appl. Meteorol. 11 (7), 1063–1070.
Seyfried, M.S., Hanson, C.L., Murdock, A.D., Van Vactor, S., 2001. Long-term lysimeter database, Reynolds Creek Experimental Watershed, Idaho, United States. Water Resour. Res. 37 (11), 2853–2856.
Stagnitti, F., Parlange, J.Y., Rose, C.W., 1989. Hydrology of a small wet catchment. Hydrol. Process. 3, 137–150.
Summer, D.M., 1996. Evapotranspiration from successional vegetation in a deforested area of the Lake Wales Ridge, Florida. US Geological Survey, Water-Resources Investigations Report 96-4244. Unpublished result, 38 pp.
Tanner, C.B., 1967. Measurement of evapotranspiration. In: Hagan, R.M., Haise, H.R., Edminster, T.W. (Eds.), Irrigation of Agricultural Lands. American Society of Agronomists, Madison, Wisconsin, pp. 534–741.
Thomas, A., 2000. Spatial and temporal characteristics of potential evapotranspiration trends over China. Int. J. Climatol. 20, 381–396.
Unland, H.E., Arain, A.M., Harlow, C., Houser, P.R., Garatuza-Payan, J., Scott, P., Sen, O.L., Shuttleworth, W.J., 1998. Evaporation from a riparian system in a semi-arid environment. Hydrol. Process. 12, 527–542.
Van Seters, T.E., Price, J.S., 2001. The impact of peat harvesting and natural regeneration on the water balance of an abandoned cutover bog, Quebec. Hydrol. Process. 15 (2), 233–248.
Wallace, J.S., Holwill, C.J., 1997. Soil evaporation from tiger-bush in southwest Niger. J. Hydrol. 188/189, 426–442.
Wilson, K.B., Baldocchi, D.D., 2000. Seasonal and interannual variability of energy fluxes over a broadleaved temperate deciduous forest in North America. Agric. For. Meteorol. 100, 1–18.
Wynn, T.M., Mostaghimi, S., Burger, J.A., Harpold, A.A., Henderson, M.B., Henry, L.-A., 2004. Variation in root density along stream banks. J. Environ. Qual. 33, 2030–2039.
Wynn, T.M., Mostaghimi, S., Harpold, A.A., Henderson, M.B., Henry, L.-A., 2003. Variation in root density along stream banks. American Society of Agricultural and Biological Engineers. Paper No. 032131.