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Partitioning evapotranspiration in semiarid grassland and shrubland ecosystems using time series of soil surface temperature

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ABSTRACT

Information about the ratio of transpiration (T) to total evapotranspiration (T/ET) is related to critical global change concerns, including shrub encroachment and non-native species invasion. In this study, a new approach was developed to partition measurements of ET into daily evaporation (E_D) and daily transpiration (T_D) in a semiarid watershed based on the low-cost addition of an infrared thermometer and soil moisture sensors to existing eddy covariance and Bowen ratio systems. The difference between the mid-afternoon and pre-dawn soil surface temperature (Δt) was used to identify days when E_D approached a seasonal minimum (E_{Dmin}) and thus, $T_D \approx ET_D - E_{Dmin}$. For other days, an empirical approach was used to partition ET_D into E_D and T_D based on volumetric soil moisture. The method was tested using Bowen ratio estimates of ET and continuous measurements of surface temperature with an infrared thermometer (IRT) at a grassland and shrubland site within the Walnut Gulch Experimental Watershed in southeast Arizona USA in years 2004–2006. Validation was based on a second dataset of Bowen ratio, IRT and shrub sap-flow measurements in 2003. Results showed that reasonable estimates of T_D were obtained for a multi-year period with ease of operation and minimal cost. Estimates of T_D and E_D were summed over the study period when plants were actively transpiring for years 2004, 2005 and 2006 to estimate totals over the study period, T_S and E_S, respectively. Preliminary analysis suggests that the accuracy of T_S estimates was 7% of the total measured sum and the precision of T_S estimates was about 4%. For this study period, T_S was related strongly to ET_S, with a slope of 0.79 for the grass-dominated site and 0.64 for the shrub-dominated site for the 3 years. Thus, for these sites during the study period in these years, the T_S/ET_S was higher for the grass-dominated site than for the shrub-dominated site, and did not vary systematically with variation in amounts and timing of precipitation. The Δt-based partitioning method has potential for international application in other well-instrumented ecosystems but will need to be tested for application when evaporation is limited by energy rather than water.

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1. Introduction

Changes in the proportion of water lost as transpiration (*T*) versus evaporation (*E*) relative to the total sum of *T* and *E* (ET) affects ecological, biogeochemical and hydrological cycles at multiple time and space scales. Information about the *T*/ET ratio has applications to critical global change concerns, including shrub encroachment, desertification, non-native species invasion, plant decomposition and soil erosion (e.g., Reynolds et al., 2000). For example, there is evidence that woody plant encroachment in arid regions may not impact the total ecosystem ET (Kurc and Small, 2004; Dugas et al., 1996; Phillips, 1992), yet it can alter the relative contributions of soil *E* and plant *T* to ET (Reynolds et al., 2000). In turn, these shifts in *E* versus *T* related to vegetation change can impact net ecosystem production and carbon cycling (Scott et al., 2006b). This has important implications for resource management strategies and other surface manipulations in dryland ecosystems associated with intensification of land use and climate change (Loik et al., 2004). Huxman et al. (2005) identified the partitioning of ET as one of the most important ecohydrological challenges in understanding vegetation dynamics in drylands.

Nonetheless, the basic understanding and prediction of *T*/ET is limited, due largely to its very complex interaction with storm size, storm intensity, storm timing, number of storms, length of interstorm periods, total seasonal precipitation, ambient meteorological conditions, plant rooting zone, soil texture, land cover and land use (e.g., Loik et al., 2004; Fernandez-Illescas et al., 2001; Bhark and Small, 2003). Field experiments designed to study *T*/ET trends at grass- and shrub-dominated sites in semiarid regions have been limited to individual precipitation events (e.g., Yepez et al., 2005) or a single growing season (e.g., Scott et al., 2006b). Reynolds et al. (2000) summarized the results of multiple studies in aridlands and reported the very different measurements of *T*/ET, varying from as little as 7% to as much as 80% in arid and semiarid ecosystems in the American Southwest (Arizona, Nevada and New Mexico). Internationally, reported values of *T*/ET in aridlands have ranged from 0.18 for grapevines and 0.34 for a ryegrass/clover mixture in Australia (Yunusa et al., 1997), 0.81 for irrigated wine and table grapes in Brazil (Teixeira et al., 2007), and about 0.45 for irrigated cowpeas in Iran (Sepaskhah and Ilampour, 1995). These studies were conducted with different methodologies at different locations over relatively short time periods. Model simulations of *T*/ET are useful for evaluating trends and relative variations over longer periods, however the accuracy of the results varies (Lane et al., 1983; Laio et al., 2001). Guswa et al. (2002) reported that ET estimated with a simple bucket model versus a more complex model at one location differed by 50% and the *T*/ET ratio varied from 55% to 67% over a season.

In semiarid regions, research has generally focused on determining the precipitation patterns and site conditions that result in a high *T*/ET ratio. There is general consensus that *E* increases with percent bare soil and *E* is related to surface soil moisture (e.g., Kurc and Small, 2004; Scott et al., 2006b). Further, total *E* over the North American monsoon period has been related to the number of storms and the length of the interstorm periods (Loik et al., 2004). In water-limited environments, models suggest that ET increases with root-zone soil moisture; *T* increases with plant canopy leaf area index (LAI); and *T* is higher in plant communities with deep roots because the plants can access more water (Porporato et al., 2001; Donohue et al., 2007). The relation between precipitation patterns and *T*/ET is less clear. Studies have shown that *T*/ET decreases with total growing-season precipitation (Loik et al., 2004; Reynolds et al., 2000); small, frequent precipitation events favor *T* over *E* (Loik et al., 2004; Monteny et al., 1997); *T*/ET is greater after large storms when deeper soil is wetted (Kurc and Small, 2007); and winter precipitation impacts total summer *T* in shrub-dominated sites (Reynolds et al., 2000). Apparently, *T*/ET depends on annual and seasonal precipitation and precipitation patterns, available energy (or potential evaporation) and plant growth form, density and LAI. Loik et al. (2004) concluded that *T*/ET must be evaluated with regard to seasonal rather than individual precipitation events, though they also found that seasonal precipitation patterns could not be simply categorized for the arid and semiarid western USA. This is the challenge that inspired the multi-year research reported here.

Multi-year studies of *T*/ET must first overcome the hurdle of measuring *T* or *E* over long periods at the ecosystem scale. The conventional approach for determining *T*/ET is to measure ET with Bowen ratio or eddy covariance methods, and then partition *T* or *E* using additional *in situ* instrumentation. Ecosystem-scale *T* has been estimated using sap-flow methods (Scott et al., 2006a) or isotope methods (Yepez et al., 2005), which are expensive and difficult to deploy and maintain in both time and space. *E* has been measured with chambers (Stannard and Weltz, 2006) and microlysimeters (Boast and Robertson, 1982; Dugas et al., 1996); however, these can also be expensive to deploy and cannot easily capture the ecosystem-scale dynamics of evaporation. As a result, there are few experimental studies of *T*/ET in drylands and few multi-year measurements to quantify the sources of variability in *T*/ET.

The basic goal of this study was to determine an operational approach for partitioning *E* and *T* from ET measurements over long time periods with ease of operation and minimal cost. A new approach is proposed to partition measurements of daily ET (ET_D) into daily evaporation (E_D) and daily transpiration (T_D) in a semiarid watershed based on the low-cost addition of an infrared thermometer and soil moisture sensors to existing eddy covariance and Bowen ratio systems. The difference between the mid-afternoon and pre-dawn soil surface temperature was used to identify days when E_D approached a seasonal minimum (E_{Dmin}) and thus, $T_D \approx ET_D - E_{Dmin}$. For other days, an empirical approach was used to partition ET_D into E_D and T_D based on upper-layer soil moisture. Instrumentation for these measurements can be maintained in place continuously for years, as demonstrated in this and other studies.

This paper describes the approach and provides validation with *in situ* measurements of T_D based on sap-flow gauges in a shrub-dominated site in 2003. The approach was applied over 3-month study periods (August to October) during the North American monsoon season in 2004, 2005 and 2006 at grass- and shrub-dominated sites in southeast Arizona USA. Results offer an insight into the hydrologic impact of woody plant encroachment in semiarid regions. Assumptions, input requirements and sources of error are discussed to guide application of the approach at other locations.

2. Study site, materials and methods

The study was conducted in the USDA Agricultural Research Service (ARS) Walnut Gulch Experimental Watershed (WGEW) located southeast of Tucson, Arizona, USA, with a semiarid climate characterized by cool, dry winters and warm, wet summers (Renard et al., 2008). The 30-year mean annual precipitation is about 350 mm (Goodrich et al., 2008) and the air temperature can range from a minimum of $-5\,^{\circ}$C to a maximum of over 40 °C (Keefer et al., 2008). Measurements were made at a grass-dominated site (Kendall) and a shrub-dominated site (Lucky Hills). The Kendall site is dominated by herbaceous vegetation, predominately grama grasses (*Bouteloua* sp.) and a variety of annual forbs. The soil at the site is dominately Stronghold (coarse-loamy, mixed, thermic Ustollic Calciorthids) with slopes ranging from 4% to 9%. Only 9 km to the west, the Lucky Hills site is located within a Chihuahuan shrub plant community dominated by creosotebush (*Larrea tridentada*) whitethorn Acacia (*Acacia constricta*) mariola (*Parthenium incanum*) and tarbush (*Flourensia Cernua*). The soil at this site is Luckyhills series (coarse-loamy, mixed, thermic Ustochreptic Calciorthids) with 3–8% slopes. The total percent vegetation cover during the study period in 2004 and 2005 was estimated to be 47% at Kendall and 40% at Lucky Hills (King et al., 2008; Skirvin et al., 2008) reflecting the slightly greater area of bare soil at the shrub-dominated Lucky Hills site. At Kendall in 2006, the relative dominance of grama grasses decreased and annual forbs increased, but the percent vegetation cover remained similar to 2004 and 2005 levels.

Over the past decades, Kendall and Lucky Hills have been instrumented with automated sensors to measure precipitation, runoff and sediment loss (symbol list provided in Table 1). As part of the Monsoon'90 Experiment, a meteorological station was added to both sites to measure basic atmospheric conditions such as air temperature, wind speed, relative humidity and solar radiation (Kustas and Goodrich, 1994). As part of Agriflux Network, automated instrumentation was installed to measure water and CO_2 flux using the Bowen ratio method to determine available energy and evapotranspiration (Svejcar et al., 1997). As part of the Soil Moisture Experiments 2004 (SMEX04) and North American Monsoon Experiment (NAME) automated soil moisture sensors and thermistors were installed at depths of 5, 15 and 30 cm and infrared thermometers were deployed to measure surface soil temperature (Jackson et al., 2007).

At each site, precipitation and runoff have been recorded from 1965 to the present (Stone et al., 2008; Goodrich et al., 2008). Since 1990, meteorological data (including incoming solar radiation, soil heat flux, and surface albedo) have been measured at 5- and/or 20-min intervals at Kendall and Lucky Hills (Keefer et al., 2008). Since 2004, soil surface temperature has been measured with an Apogee IRTS-P[1] infrared thermometer (IRT) at 20-min intervals. Potential ET (ET_P) was computed for each time step using the Penman-Monteith equation (Allen, 1986) for WGEW (per Moran et al., 1994, Appendix 1) and summed to a daily total (ET_{PD}) for each day in the study period.

Table 1 – Symbol list, ordered primarily by measurement type

E = evaporation
- E_D = daily E (mm day^{-1})
- E_S = E summed over the study period (mm season^{-1})
- E_{Dmin} = seasonal minimum daily E (mm day^{-1})

ET = evapotranspiration
- ET_D = daily ET (mm day^{-1})
- ET_S = ET summed over the study period (mm season^{-1})
- ET_{PD} = daily potential ET (mm day^{-1})

T = transpiration
- T_D = daily T (mm day^{-1})
- T_S = T_D summed over the study period (mm season^{-1})
- T_{S31E} = T_D estimates summed over a 31-day period in 2003 (mm season^{-1})
- T_{S31M} = T_D measurements summed over a 31-day period in 2003 (mm season^{-1})

θ = volumetric soil moisture (cm^3 cm^{-3})
- $\theta_{@5}$, $\theta_{@15}$, and $\theta_{@30}$ = θ measured at 5, 15 and 30 cm, respectively (cm^3 cm^{-3})
- θ_{15} = upper-layer soil moisture to about 15 cm depth defined by Eq. (1) (cm^3 cm^{-3})

t_S = soil surface temperature measured with a down-looking infrared thermometer (°C)
- t_{Smin} = minimum t_S between 3:00 and 5:00 a.m. (°C)
- t_{Smax} = maximum t_S between 1:00 and 3:00 p.m. (°C)
- t_{Ssim} = simulated t_S defined by Eq. (5) (°C)
- Δt = difference between t_{Smax} and t_{Smin} (°C)
- $\Delta t'$ = detrended Δt values (°C)

P = soil thermal inertia (J m^{-2} K^{-1} s$^{-1/2}$)
- k = thermal conductivity (W m^{-1} K^{-1})
- ρ = dry bulk density (kg m^{-3})
- c = specific heat capacity (J kg^{-1} K^{-1})
- ATI = apparent thermal inertia defined by Eq. (4) (°C)

α = surface albedo (unitless)
C_v = volumetric heat capacity of air (J °C^{-1} m^{-3})
G = soil heat flux density (W m^{-2})
r_a = aerodynamic resistance (s m^{-1})
R_n = net radiant flux density (W m^{-2})
$R_{S\downarrow}$ = incoming solar radiation (W m^{-2})
t_a = air temperature (°C)
VPD = vapor pressure deficit (kPa)
u = wind speed (m s^{-1})

Since 2004, volumetric soil moisture (θ) has been measured at three depths (5, 15 and 30 cm) with Stevens Water Hydra Probe sensors at 20-min intervals (Keefer et al., 2008). These sensors have an integrated thermistor to measure soil temperature at the same depths and time intervals. Soil temperature was also measured at 1-, 2-, and 6-cm depths with thermocouples at 20-min intervals. Soil moisture measured at 5 cm ($\theta_{@5}$) was assumed to characterize the surface soil moisture from 0 to 5 cm (θ_5). Upper-layer soil moisture to about 15 cm depth (θ_{15}) was computed as

$$\theta_{15} = \frac{\theta_{@5} + \theta_{@15}}{2}. \tag{1}$$

Similarly, soil moisture to about 30 cm depth (θ_{30}) was computed as the average of soil moisture measurements at 5, 15 and 30 cm ($\theta_{@5}$, $\theta_{@15}$, and $\theta_{@30}$, respectively).

[1] Use of trade names in this paper is for information purposes only and does not constitute an endorsement by the USDA-ARS.

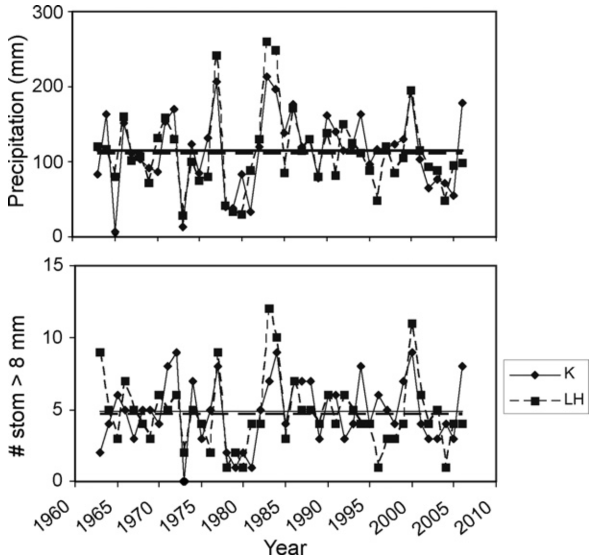

Fig. 1 – The precipitation and storm count for storms with precipitation greater than 8 mm (#storm > 8 mm) for the study period (DOY 215–285) at the grass-dominated Kendall (K) and shrub-dominated Lucky Hills (LH) sites, relative to the mean precipitation and storm count for 1963–2006.

These sites were also equipped with flux stations to measure evapotranspiration using the Bowen ratio technique at 20-min intervals from 1997 to 2007 (Emmerich, 2007; Emmerich and Verdugo, 2008b). The Bowen ratio systems were placed in locations with a fetch of 200+ m in all directions at the two sites. CO_2 flux measurements from 1997 to the present at both Kendall and Lucky Hills showed that during the period from day of year (DOY) 215–285 the plants were most likely to be transpiring when moisture was available (Emmerich and Verdugo, 2008a). Thus, we defined the study period for this analysis from DOY 215 to 285 (1 August to 10 October in years that were not leap years) during the North American monsoon season. The precipitation patterns of study periods in years 2004–2006 at the Kendall and Lucky Hills sites, relative to the entire precipitation record, are presented in Fig. 1 and Table 2. The mean annual precipitation from 1963 to 2006 was 340 mm at Kendall and 320 mm at Lucky Hills.

During the North American monsoon in 2005 (from DOY 198 to 260) at Kendall, daily soil evaporation was measured manually from a network of 20 microlysimeters on days between precipitation events (Green, 2006). Microlysimeters of 76-mm diameter and 30-cm depth were installed in a cross-shaped pattern centered on the Bowen ratio system over an area of 60 m × 60 m. An identical network of microlysimeters was deployed at Lucky Hills in 2007 and measurements of daily soil evaporation were made from DOY 253 to 267 at the end of the North American monsoon.

During the North American monsoon in 2003, shrub transpiration was measured at the Lucky Hills site every 30 min using the constant heat balance sap-flow technique (Scott et al., 2006a). Stems of 16 shrubs were instrumented to capture the diurnal dynamics of canopy water loss within the footprint of the Bowen ratio system. This shorter, but more comprehensive, dataset was used to validate the approach for partitioning ET developed here.

3. Approach for partitioning ET

The approach for partitioning ET is a unique and critical part of this study. It is based on the assumption that conventional eddy covariance or Bowen ratio instrumentation is in place at a site, with coincident measurements of soil surface temperature and upper-layer soil moisture, making measurements throughout the day at the commonly used 20- or 30-min time interval. The difference between the mid-afternoon and pre-dawn soil surface temperature (Δt) was used to identify

Table 2 – Precipitation patterns at Kendall and Lucky Hills for years 2002–2006, where study period refers to DOY 215–285, *winter* covers the period DOY 1–90, *annual* covers the period from 1 January to 31 December, #st > 2 mm and #st > 8 mm refer to the number of storms during the study period with infiltration >2 mm and >8 mm (respectively) and %dp < 10 day refers to the percent of dry periods (periods with no precipitation) that were shorter than 10 days during the study period

Year	#st > 2 mm during study period	#st > 8 mm during study period	%dp < 10 day during study period	Winter precipitation (mm)	Precipitation during study period (mm)	Annual precipitation (mm)
Kendall						
2002	12	4	73	27	92	236
2003	8	2	63	42	87	200
2004	10	5	78	104	101	295
2005	8	3	71	65	61	162
2006	19	8	89	7	195	274
Lucky Hills						
2002	11	5	73	18	111	247
2003	9	4	60	40	105	246
2004	10	1	63	104	65	219
2005	10	4	80	69	107	223
2006	19	4	94	6	112	189

Antecedent precipitation data for 2002 was included to provide readers with information that will drive initial soil moisture conditions during the 2003–2006 study.

days when E_D approached a seasonal minimum based on the link between Δt and soil thermal inertia.

By definition, soil thermal inertia (P) represents the ability of soil to conduct and store heat, where

$$P = (k\rho c)^{1/2}\ [\mathrm{J\,m^{-2}\,K^{-1}\,s^{-1/2}}], \tag{2}$$

and k = thermal conductivity [W m^{-1} K^{-1}]; ρ = dry bulk density [kg m^{-3}]; and c = specific heat capacity [J kg^{-1} K^{-1}] where volumetric heat capacity is the product of the last two terms. Like P, Δt also represents the capacity to store heat and the resistance of soil to temperature change, where

$$\Delta t = (t_{Smax} - t_{Smin})\ [^\circ\mathrm{C}]. \tag{3}$$

The terms t_{Smax} and t_{Smin} represent soil surface temperatures measured with a down-looking infrared thermometer (IRT) selected to be the maximum between 1:00–3:00 p.m. and the minimum between 3:00–5:00 a.m., respectively. In early studies, Δt was linked theoretically to P resulting in the apparent thermal inertia (ATI) derived by Price (1977) where

$$\mathrm{ATI} = \frac{1-\alpha}{\Delta t}\ [^\circ\mathrm{C}], \tag{4}$$

and α is the surface albedo. ATI was related to regional soil moisture (Kahle, 1987; Pratt and Ellyett, 1979) but it was not easily interpreted over a heterogeneous terrain and determined to be of limited utility (Price, 1985). The use of surface temperature for deriving P was revitalized by Xue and Cracknell (1995) and Cracknell and Xue (1996a) with a simple model to determine P (as distinct from ATI) based on the time of maximum temperature in the daytime. This work confirmed that P was directly proportional to Δt and launched renewed interest in the use of Δt for mapping soil moisture and heat flux (e.g., Cracknell and Xue, 1996b; Verhoef, 2004; Cai et al., 2007; Sobrino et al., 1998; Verstraeten et al., 2006).

It is demonstrated herein that when Δt reached a seasonal maximum, E_D approached a minimum value. With Δt at the seasonal maximum, ET_D can be measured and dates for which E_D approaches the seasonal minimum ($E_D \rightarrow E_{Dmin}$) can be identified. For these days, T_D can then be assumed equal to $ET_D - E_{Dmin}$. It is known the E_D rarely equals zero, but rather approaches a minimal value that is substantially smaller than T_D during the growing season. For example, Albertson et al. (1995) reported that for dry conditions at Owens Lake, California, USA, E_D can still be 5% of net radiation. At WGEW, E_{Dmin} was reported to average 0.34 mm day^{-1} during the North American monsoon season based on microlysimeter measurements; this corresponded to a E_D/ET_{PD} ratio of less than 0.1 (Green, 2006).

For days when E_D is not determined to be at the seasonal minimum, an empirical supplement is proposed to determine T_D. Using Δt-derived estimates of T_D for days when $E_D \approx E_{Dmin}$ in 2004–2006, a relation was determined between T_D and θ_{15} at the grass- and shrub-dominated sites. This relation was used to estimate T_D based on the measurement of θ_{15} and the residual $ET_D - T_D$ difference was attributed to E_D (explained in Section 3.2). With this set of measurements and equations, it was possible to partition measurements of ET_D into T_D and E_D for all days in the study period.

3.1. *Computing, detrending and thresholding Δt with time-series t_S measurements*

The approach for computing, detrending and thresholding Δt was determined with a time series of t_S measurements made every 20 min during the study period (DOY 215–285) from years 2004–2006. Δt can be computed by simply taking the difference between the maximum and minimum soil surface temperature measured within the hours of 1:00–3:00 p.m. and 3:00–5:00 a.m., respectively (Eq. (3)). However, since time-series t_S measurements were available every 20 min, we used the entire diurnal dataset to improve estimates of t_{Smax} for days with partial cloud interference. The assumption was that for clear-sky days without precipitation, the 20-min t_S measurements can be fit with a sine curve of the form $A \sin(Bx + C)$ where x is the time of day between sunrise to sunset and A, B and C are coefficients defining the shape and magnitude of the sine curve. To discard t_S measurements that had been attenuated by cloudy conditions, a data point was eliminated if it did not represent a convex curve relative to its 20-min neighbors in the diurnal time series. The remaining data points were used to fit a sine curve using a constrained non-linear optimization (CNO) procedure on a root-mean-squared (RMS) error function. The constraint is that the actual data points must lie on or beneath the extrapolated curve. As a result, if the measurements are accurate and the CNO procedure is implemented correctly, there is no possibility

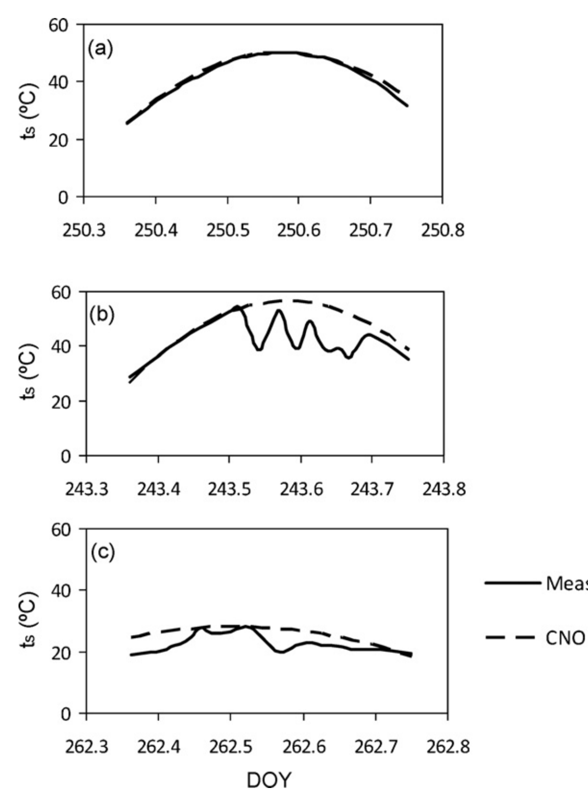

Fig. 2 – Examples of the constrained non-linear optimization (CNO) procedure for estimating daytime t_S on (a) clear-sky days, (b) partially cloudy days and (c) very cloudy days in 2004, where the solid line represents the on-site measurement and the dashed line represents the sine curve fit to filtered measurements.

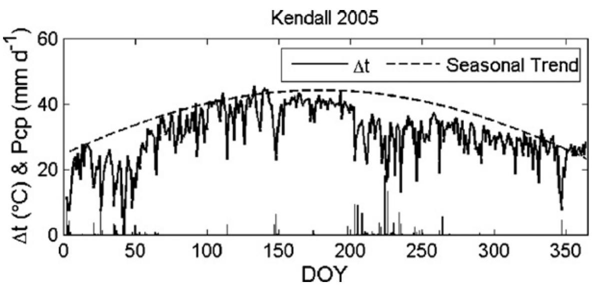

Fig. 3 – Example of the approach for detrending Δt at Kendall 2005, where the solid line represents the on-site measurements of Δt and the dashed line represents the sine curve fit to the highest Δt values (determined for 20-day intervals over the year). The difference between the measurement and the seasonal trend was used to produce detrended Δt values (termed Δt′) over the year. The bars represented daily precipitation (mm day^{-1}).

for the extrapolated curve to obtain a higher temperature than the highest possible daily temperature (Fig. 2a and b). On the other hand, it is unavoidable that the CNO method will produce an extrapolated curve that does not obtain the highest possible daily temperature when there is not enough cloud-free data to properly fit a curve (Fig. 2c). This is acceptable since these days would not have been useful for this approach, with either the original data or this CNO procedure. Using the CNO procedure to estimate t_{Smax}, 1–7 more days per study period per year were identified for which $E_D \approx E_{Dmin}$.

Values of Δt computed with Eq. (3) followed a seasonal trend in which higher values were obtained during the summer when solar radiation was at a maximum (explained by McVicar et al., 2007). To extract the days when $E_D \approx E_{Dmin}$ using a single threshold, it was first necessary to detrend the annual Δt time series. This was accomplished by making an RMS fit to a sine curve over the high Δt values determined for 20-day intervals over the year (Fig. 3). The Δt values tracked the sine curve well during the hot dry season (DOY 100–180) when soil moisture was low, and fell below the curve during the hot wet season (DOY 200–300) when soil moisture was variable. Care should be taken with detrending since the partitioning results depend largely on the method chosen to remove the seasonal trend. The detrended Δt values, termed Δt′, are the basis for the partitioning results reported in Section 4.

Once detrended, the next step was to set a Δt′ threshold to identify dates for which E_D approaches the seasonal minimum

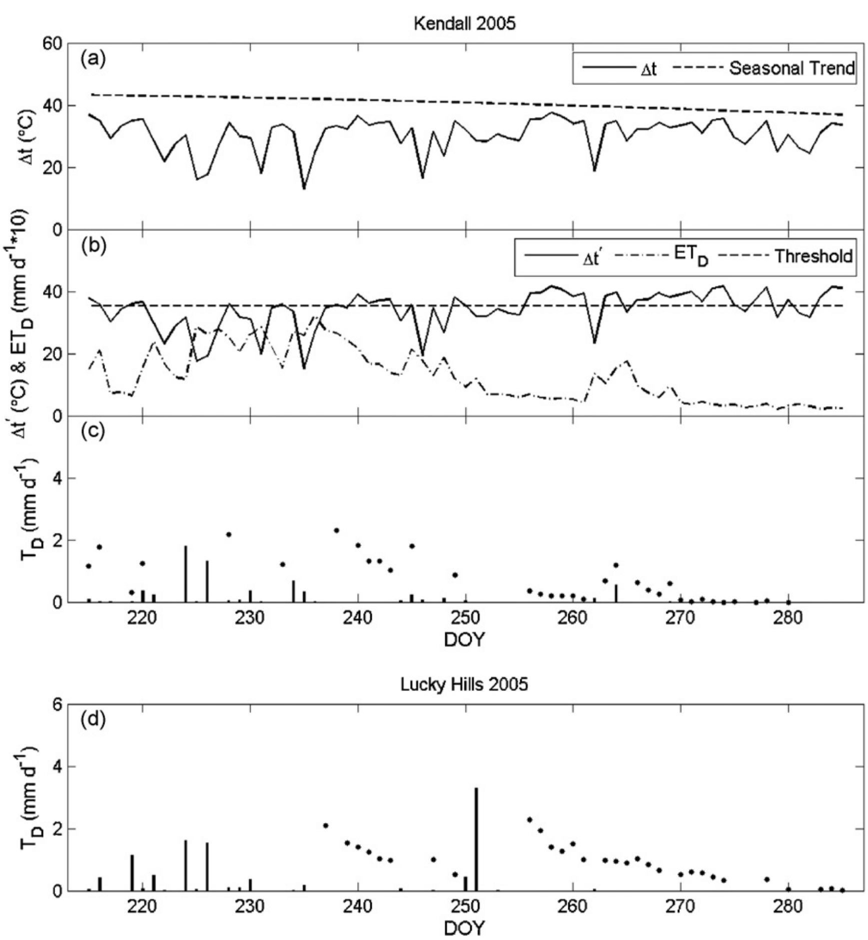

Fig. 4 – An illustration of the steps taken to partition ET using Δt at Kendall in 2005. (a) A sine curve was fit to the highest Δt values. (b) A threshold was set to select Δt′ values that were within 15% of the highest Δt′ value during the study period. (c) For dates when Δt′ exceeded the threshold, we presumed that $E_D \approx E_{Dmin}$ and $T_D \approx ET_D - E_{Dmin}$. (d) Values of T_D for Lucky Hills were derived using the same process illustrated at Kendall in (a)–(c). Bars represent daily precipitation (cm day^{-1}).

($E_D \rightarrow E_{Dmin}$). A threshold was set to select $\Delta t'$ values that were within 15% of the highest $\Delta t'$ value during the study period. That is, if the highest $\Delta t'$ value for the season was 50 °C, then the threshold was set at 42.5 °C. For dates which $\Delta t'$ exceeded that threshold, we presumed that $E_D \approx E_{Dmin}$ and $T_D \approx ET_D - E_{Dmin}$. This threshold was set by running a Monte Carlo simulation (Press et al., 1986) to compute the uncertainty of the highest $\Delta t'$ value due to reasonable variations in meteorological and surface conditions (e.g., McVicar and Jupp, 2002). For the simulation, surface temperature of a dry bare soil (°C) at midday was computed as

$$t_{Ssim} = t_a + \left[\frac{r_a(R_n - G)}{C_v}\right] \quad (5)$$

where t_{Ssim} is simulated soil surface temperature (°C), t_a is the air temperature (°C), r_a the aerodynamic resistance (s m^{-1}), R_n is the net radiant flux density (W m^{-2}), G is soil heat flux density (W m^{-2}), and C_v is the volumetric heat capacity of air (J °C^{-1} m^{-3}) (Jackson et al., 1981). The measurements used to solve Eq. (5) were incoming solar radiation ($R_{S\downarrow}$, W m^{-2}), t_a, t_{Smax}, vapor pressure deficit (VPD, kPa), wind speed (u, m s^{-1}) and albedo (α) (Moran et al., 1994) measured at midday on the day when $\Delta t'$ was highest during the study period for each year at each site. Eq. (5) was solved 10,000 times while varying the inputs randomly within specified confidence ranges of $R_{S\downarrow}$ (±10%), t_a (±2 °C), t_{Smax} (±2 °C), VPD (±10%), u (±10%) and α (±5%). This would simulate slightly varying conditions for clear-sky days when the soil was relatively dry and $E_D \approx E_{Dmin}$. The uncertainty in t_{Ssim} was determined to be approximately 15% based on two standard deviations from the mean of 10,000 simulations.

In summary, the Δt values for study periods in 2004, 2005 and 2006 were determined based on Eq. (3) where the t_{Smin} was the minimum soil surface temperature measured with the down-looking IRT between 3:00 and 5:00 a.m. The IRT measurements and the CNO procedure were used to determine the t_{Smax} between 1:00 and 3:00 p.m. All Δt values were detrended using the sine curve fit to the high values of Δt, described previously (Fig. 4a). A threshold was set to select detrended Δt values ($\Delta t'$) that were within 15% of the highest $\Delta t'$ value during the study period (Fig. 4b). For dates which $\Delta t'$ exceeded that threshold, we presumed that $E_D \approx E_{Dmin}$ and $T_D \approx ET_D - E_{Dmin}$. Thus, T_D was estimated for selected dates for the grass- and shrub-dominated sites (Kendall and Lucky Hills, respectively) (Fig. 4c and d).

3.2. Partitioning ET_D when $E_D \neq E_{Dmin}$ and/or for cloudy sky conditions

The Δt-based approach was used to estimate T_D for selected dates in 2004, 2005 and 2006, as illustrated for 2004 in Fig. 4. The relation between these estimates of T_D and upper-layer soil moisture to 15 cm depth (θ_{15}) was non-linear and statistically significant (Fig. 5). For Kendall,

$$\begin{aligned} T_D &= (33.2\theta_{15}) - 1.2\,[r^2 = 0.72,\ p < 0.01];\ \text{for}\ \theta_{15} < 0.11\,\text{cm}^3\,\text{cm}^{-3}, \\ T_D &= 2.5\,(\text{mm/day});\ \text{for}\ \theta_{15} \geq 0.11\,\text{cm}^3\,\text{cm}^{-3}, \end{aligned} \quad (6)$$

and for Lucky Hills,

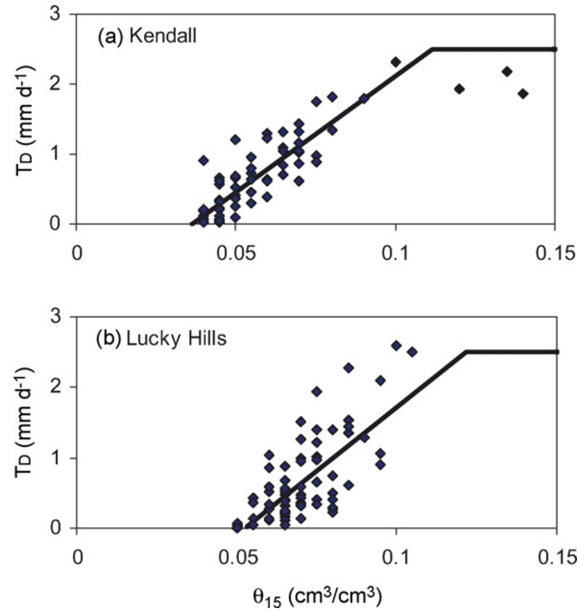

Fig. 5 – The relation between Δt-derived estimates of T_D for 2004–2006 during the study period (DOY 215–285) and upper-layer soil moisture to 15 cm depth (θ_{15}) for Kendall and Lucky Hills.

$$\begin{aligned} T_D &= (36.1\theta_{15}) - 1.9\,[r^2 = 0.50,\ p < 0.01];\ \text{for}\ \theta_{15} < 0.12\,\text{cm}^3\,\text{cm}^{-3}, \\ T_D &= 2.5\,(\text{mm/day});\ \text{for}\ \theta_{15} = 0.12\,\text{cm}^3\,\text{cm}^{-3}. \end{aligned} \quad (7)$$

For the T_D and θ_{15} relation (Fig. 5) an upper limit was set at T_D = 2.5 mm day^{-1}, above which T_D was a constant unrelated to increasing root zone soil moisture (as suggested by Laio et al. (2001) and supported by Kurc and Small (2004)). The 2.5 mm day^{-1} ceiling was set based on season-long estimates of T_D at Kendall and Lucky Hills that did not exceed 2.5 mm day^{-1} (see Fig. 5 and results by Green, 2006). This relation between T_D and θ_{15} was used to supplement the Δt-based approach by determining T_D on days when E_D was unknown. We found weaker relations between T_D and soil

Table 3 – Correlation coefficients (r^2) between daily transpiration (T_D) and soil moisture measured at 5 cm (θ_5) and within the upper layer to 15 and 30 cm depths (θ_{15} and θ_{30}) at Kendall and Lucky Hills

	Correlation coefficients (r^2)			
	T_D		T_D/ET_{PD}	
	Kendall	Lucky Hills	Kendall	Lucky Hills
θ_5	0.47	0.15	0.47	0.12
θ_{15}	0.72	0.50	0.71	0.50
θ_{30}	0.33	0.43	0.37	0.45

Estimates of T_D were divided by total daily potential ET (ET_{PD}) and again related to soil moisture to confirm that plant transpiration was predominately water limited, not energy limited, during the study period (DOY 215–285).

moisture measured at 5, 15 and 30 cm or the average soil moisture to 30 cm depth (θ_{30}) (Table 3).

To account for differences in atmospheric conditions (particularly available energy) in development of this relation, estimates of T_D were divided by ET_{PD} and again related to θ_{15}. The correlation coefficients did not show substantial improvement over the coefficients for the relation between T_D and θ_{15} (Table 3) implying that plant transpiration was water limited, not energy limited, for the vast majority of days during the study period. This is a reasonable result since T_D and ET_{PD} do not covary at our semiarid sites during the North American monsoon season, resulting in non-significant correlation coefficients of 0.01 at both Kendall and Lucky Hills. Similar results were reported for grass- and shrub-dominated sites in New Mexico by Kurc and Small (2004). The ratio of the sum of T_D to the sum of ET_{PD} over DOY 215 to 285 was 0.28, 0.25 and 0.54 at Kendall and 0.20, 0.43 and 0.55 at Lucky Hills for years 2004, 2005 and 2006, respectively.

4. Results and discussion

Analysis was conducted with datasets extracted for the study period (DOY 215–285) from years 2004–2006. The first two subsections present results and validation of the empirical partitioning approach. The final subsection presents results of analysis of these estimates of T_D, E_D and their sums over the study period (T_S and E_S, respectively) in 2004, 2005 and 2006 for the grass- and shrub-dominated sites.

4.1. *Relation between ET_D, E_D and $\Delta t'$*

As discussed earlier, $\Delta t'$ is theoretically related to both surface and atmospheric conditions. This sensitivity is illustrated by the response of $\Delta t'$ to a variety of surface and atmospheric conditions associated with a spring storm that occurred at Kendall in 2005 before the summer vegetation growth (Fig. 6a). A precipitation event on DOY 147 and 148 resulted in a dramatic decrease in $\Delta t'$ associated with an increase in ET_D (due to increased soil moisture) and an associated decrease in available solar energy (due to cloudiness). For the clear-sky days that followed the storm event (DOY 149–152) $\Delta t'$ steadily increased as ET_D decreased, finally reaching a value similar to that before the storm. However, cloudy conditions on the following day (DOY 153) resulted in another dramatic decrease in $\Delta t'$ without an associated increase in soil moisture. Surface temperature is related to air temperature and solar radiation, as expressed in Eq. (5) and both daily maximum and minimum air temperatures are a function of solar radiation dynamics (Bristow and Campbell, 1984). This demonstrates the difficulty in interpretation of $\Delta t'$, and introduces the rationale behind the approach used here.

For a winter storm in 2004 at Kendall (when transpiration was assumed to be near zero) the $\Delta t'$ related well with ET_D (Fig. 6b). Generally, the $\Delta t'$ increased as ET_D decreased, and both measures returned to their pre-storm values within the same time period (about 7 days after the storm). For the 2004 summer storm (Fig. 6c) the $\Delta t'$ post-storm recovery corresponded to a steep decline in ET_D, followed by a more gradual

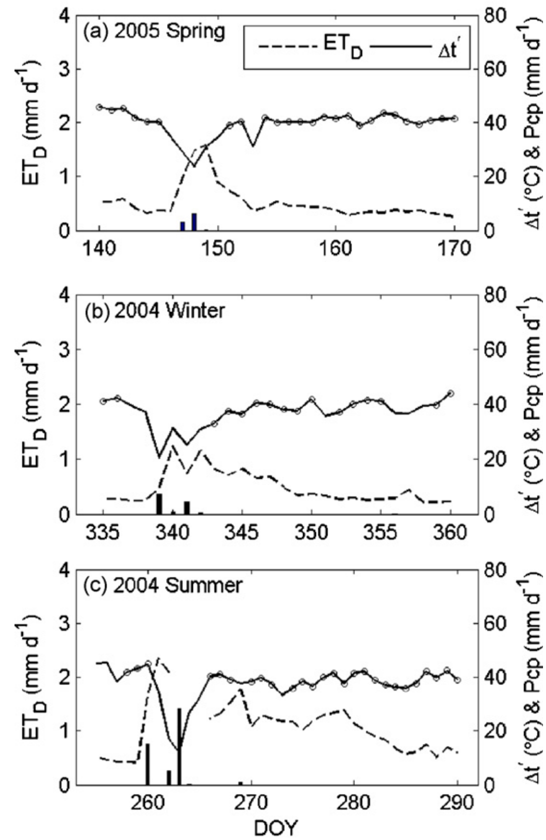

Fig. 6 – Comparison of $\Delta t'$ derived from IRT measurements (°C) with daily ET measured with the Bowen ratio method (mm day^{-1}) for (a) spring, (b) winter and (c) summer storms at Kendall in 2004 and 2005, followed by a series of predominantly cloudfree days. The break in ET measurements from DOY 263–265 was due to instrument failure. Similar results were found (though not shown here) for Lucky Hills. Circles indicate the cloudfree days and bars represent daily precipitation (mm day^{-1}).

decline in ET_D. The $\Delta t'$ returned to its pre-storm value about 4 days after the storm. This trend was confirmed by the $\Delta t'$ and E_D measurements made at Lucky Hills in 2003 (Fig. 7). For two small summer storms, variation (decrease and recovery) in $\Delta t'$ corresponded well to the measured increase and subsequent decrease in E_D. The key to this approach is that $\Delta t'$ reaches a stable value on the day that E_D approaches E_{Dmin}. Based on these observations, we postulate that the highest $\Delta t'$ values were associated with cloudfree days when $E_D \rightarrow E_{Dmin}$.

4.2. *Validation of the empirical partitioning approach*

The accuracy of the T_D values estimated using the Δt-based approach and the θ_{15}-based supplement was determined using the measurements of T_D from a study with sap-flow sensors installed at Lucky Hills in 2003 (Scott et al., 2006a). The same series of steps illustrated in Fig. 4 for Kendall, 2004 were followed to estimate T_D for clear-sky days when E_D approached E_{Dmin} using $\Delta t'$ at Lucky Hills in 2003. For all other days, T_D was computed using the empirical relation with

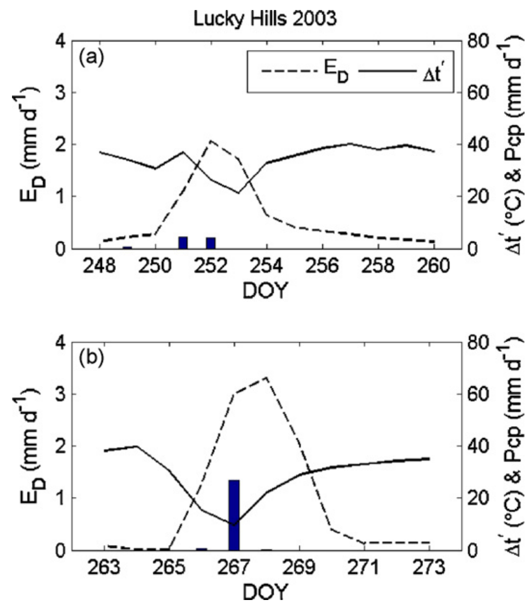

Fig. 7 – Comparison of $\Delta t'$ with daily E (E_D) for two storms in 2003 at Lucky Hills, when E_D was determined from the difference between ET_D (using Bowen ratio) and T_D (using sap-flow technique). Bars represent daily precipitation (mm day^{-1}).

θ_{15} (Eqs. (6) and (7)). The root mean squared error (RMSE) of T_D estimated with $\Delta t'$ and T_D measured with the sap-flow technique was 0.29 mm day^{-1} (illustrated with asterisks in Fig. 8). The correlation between T_D estimated with the θ_{15}-based supplement was also reasonable (illustrated with solid circles in Fig. 8, RMSE = 0.26 mm day^{-1}). The worst result (absolute difference of 0.9 mm day^{-1}) was obtained immediately after a large storm on DOY 239. Apparently, when the θ_{15}-based supplement is used to estimate T_D, this approach is prone to overestimate T_D (and underestimate E_D) during the days immediately following storms.

Accuracy at the month- or season-long time scale was assessed by summing daily values of T_D or E_D over all measurement days in the study period. For the 2003 study period, T_D values were summed to produce a 31-day transpiration total for estimates and measurements (T_{S31E} = 39 and T_{S31M} = 42 mm, respectively). For this 31-day sum, the difference between T_{S31E} and T_{S31M} was 3 mm, or 7% of the total measured sum. This exercise offers an assessment of the accuracy that might be expected for sums of T_D and E_D over prolonged periods. In general, when more days during a given study period are estimated with the Δt-based approach than with the θ_{15}-based supplement, the accuracy will improve.

The precision of sums of T_D (T_S) for the 2004–2006 study periods was determined using standard error propagation methods (Taylor, 1997). The variance of T_D for any given day estimated with the Δt-based approach was determined by assuming a 95% probability that the estimated value of T_D was within 15% of the true value (the threshold selected in Fig. 4b) and using the cumulative distribution of the standardized normal distribution table to solve for the variance. The variance of T_D for any given day estimated with the θ_{15}-based supplement was computed from the error matrix derived with the least squared method for determining the coefficients in

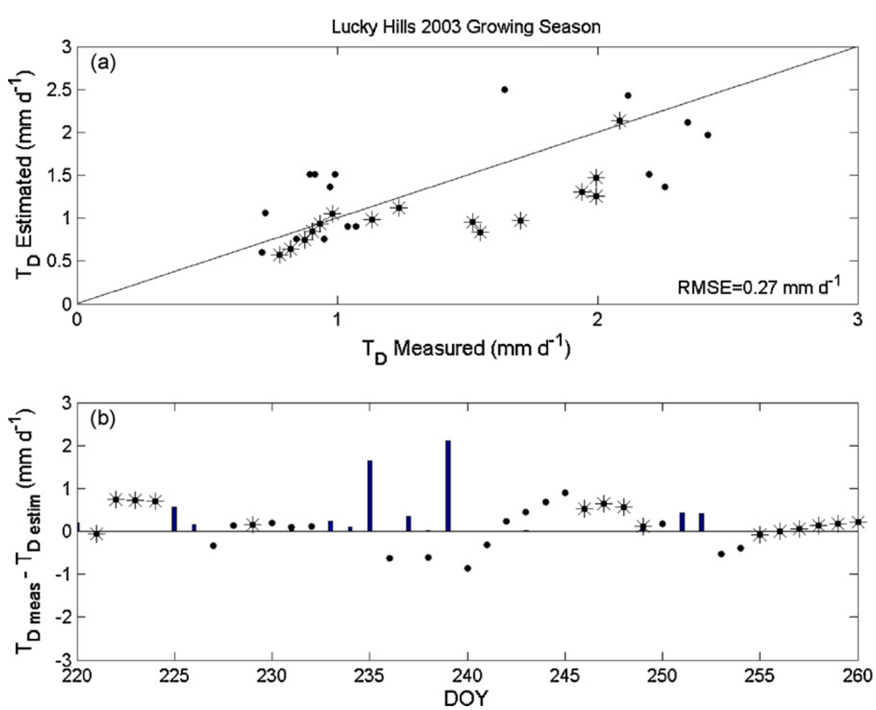

Fig. 8 – (a) Comparison of daily transpiration estimated with this approach (T_D Estimated) and daily transpiration measured with a conventional sap-flow technique (T_D Measured, from Scott et al., 2006b) and (b) the difference between measured and estimated T ($T_{Dmeas} - T_{Destim}$). The results estimated with $\Delta t'$ are illustrated with an asterisk, and results estimated with the supplemental approach are illustrated with a solid circle. The bars represented daily precipitation (cm day^{-1}).

Table 4 – Precision of the sum of T_D over the study period (T_S) presented in units of mm and as percent of the total sum (%) in 2004, 2005 and 2006 for the grass- and shrub-dominated sites, where precision was determined using standard error propagation methods

Year	Site	T_S (mm)	Precision of T_S (mm)	Precision of T_S (%)
2004	Kendall	99.8	4.0	4.0
	Lucky Hills	73.5	3.3	4.5
2005	Kendall	89.4	3.8	4.3
	Lucky Hills	113.6	5.9	5.2
2006	Kendall	162.4	6.0	3.7
	Lucky Hills	142.5	7.0	4.9

The precision for any given year and location was a function of the number of days T_D is computed with the Δt-based approach and with the θ_{15}-based supplement.

Eqs. (5) and (6) and a first-order Taylor series (Bevington and Robinson, 1992) to propagate these errors and estimate the variance. In this way, we were able to compute the variance of T_D on every day in the study period and sum these variances to determine the precision of the T_S summed over the study period. For these 3 years at these 2 sites, the precision ranged from about 3–7 mm or about 4% of T_S (Table 4). The precision for any given study period was partly a function of the number of days T_D is computed with the Δt-based approach and with the θ_{15}-based supplement.

4.3. *Trends in E_S and T_S over the study period*

Estimates of E_D and T_D for the study period were determined by combining the Δt-based partitioning approach and the θ_{15}-based supplement. Estimates of E_D and T_D were summed over the study period to determine the total evaporation and transpiration (E_S and T_S, respectively).

T_S over the 3 years was highly correlated with ET_S, resulting in T_S/ET_S values that were relatively constant over this 3-year period at WGEW. For Kendall,

$$T_S = (0.79 ET_S)\ [r^2 = 0.96,\ p < 0.01] \tag{8}$$

and for Lucky Hills,

$$T_S = (0.64 ET_S)\ [r^2 = 0.92,\ p < 0.01]. \tag{9}$$

The T_S/ET_S values, defined by the slope of the relation between T_S and ET_S over the 3-year period, were 0.79 for Kendall and 0.64 for Lucky Hills. The difference between the mean T_S/ET_S at Kendall and Lucky Hills over the 3 years was statistically significant at $p < 0.01$. Thus, for these sites during the study period in these years, the T_S/ET_S was higher for the grass-dominated site than for the shrub-dominated site. This may be related to the slightly higher vegetation cover at Kendall than at Lucky Hills. It may also be related to the different ecosystem water use efficiencies of the two plant communities (McVicar et al., 2002). Emmerich (2007) found that the grass-dominated Kendall site was 1.4–1.6 times more water use efficient than the shrub-dominated Lucky Hills site.

Since T_S/ET_S was found to be relatively constant over this study period that was characterized by distinctive storm patterns (Table 2) it followed that there were weak relations between T_S/ET_S and total infiltration (precipitation minus runoff), the number of large ($\geq$8 mm) storms, and the total number of storms ($\geq$2 mm) and total ET_S during the study period (Table 5). This statement must be immediately clarified. This conclusion is limited to the study period defined here (DOY 215–285 within the North American monsoon) and is constrained by the fact that we are not studying shrub encroachment, but rather the difference between grass- and shrub-dominated sites located 9 km apart. For this study, T_S/ET_S for two sites was independent of precipitation patterns, but significantly dependent on vegetation type (grass-dominated versus shrub-dominated).

Table 5 – Correlation coefficients (r^2) between T_S/ET_S and precipitation patterns for the study periods in 2004, 2005 and 2006 (n = 3) such as total infiltration (precipitation minus runoff), the number of large storms with infiltration ≥ 8 mm, total number of storms (with infiltration ≥ 2 mm), and total ET_S during the study period

	Kendall	Lucky Hills
Total infiltration	0.06	0.05
Number of storms with infiltration ≥ 8 mm	0.05	0.04
Number of storms with infiltration ≥ 2 mm	0.21	0.61
Total ET_S	0.14	0.01

None of the correlation coefficients were significant at $p < 0.01$.

5. Application of the empirical partitioning approach

The assumptions, input requirements and sources of error associated with this approach for partitioning ET warrant some discussion. It is first important to clarify the distinction between the Δt-based approach for partitioning ET_D and the empirical supplement. The Δt-based approach estimates E_D using primarily IRT measurements of soil surface temperature; the result is an estimate of T_D for a limited number of days. The supplement is based on using those limited T_D estimates to develop an empirical relation with another *in situ* measurement (in our case, θ_{15}). Thus, the Δt-based approach has value in itself, but the θ_{15}-based supplement is dependent on the Δt-based partitioning results. There are a set of assumptions, inputs and errors associated with the Δt-based approach and its supplement, and with application of both approaches combined.

5.1. Δt-based approach

A basic assumption associated specifically with the Δt-based approach is that there is a maximum temperature a given dry soil surface can reach under clear sky conditions driven by t_a, r_a, and C_v for the site and the season (Eq. (5)). This builds on well-accepted theory presented by Jackson et al. (1981). To account for the seasonal variations in solar radiation and the smaller day-to-day variations in wind speed and vapor pressure deficit, seasonal measurements of Δt were detrended to $\Delta t'$ values and a threshold was set to select $\Delta t'$ values that were within 15% of the highest $\Delta t'$ value during the study period. For the study presented here, we detrended the data over a 12-month period. In an informal test, we found that the dates identified when $E_D \approx E_{Dmin}$ were the same using data that were detrended for periods ranging from several months to 1 year. The detrending curves remained relatively consistent from year to year at a given site but did not afford coherent collusion between years. This generally resulted in a difference of 1–2 °C from the last day of 1 year to the first day of the next during the winter, but under some conditions, the difference could be as large as 10 °C. The diurnal temperature detrending is not a well understood practice and should be considered highly application and location specific.

Regarding the threshold set to select $\Delta t'$ values within 15% of the highest $\Delta t'$ value, this threshold may differ with location. We set a conservative threshold to encompass the uncertainty associated with the combined day-to-day variations in available energy, wind speed, vapor pressure, air temperature and the error of the IRT measurement and deployment. The threshold would likely have to be recomputed for different study periods and locations.

This approach was applied to a region (WGEW) and season (North American monsoon) for which site evaporation is generally limited by water, not available energy (Reynolds et al., 2000). The approach will have to be tested for application when evaporation is limited by energy rather than water. Under those conditions, more clouds would be present and the Δt detrending approach would be even more important. Also, in energy-limited environments, the influence of diffuse light on plant light use efficiency would increase and might deserve attention (Roderick et al., 2001). Likely the $\Delta t'$ will be associated with E_D/ET_{PD}, where ET_{PD} can be computed directly from data already measured with eddy covariance or Bowen ratio systems. Further, the relation in Fig. 5 would correlate θ_{15} with T_D/ET_{PD} to account for the sensitivity to available energy. This approach will need to be tested, and possibly refined, for application under energy-limited conditions.

There are assumptions in the Δt-based approach associated with the deployment of the IRT. First, the IRT is assumed to be pointing at a bare soil target representative of conditions within the footprint of the Bowen ratio ET_D measurement. That is, the measurement of t_S over an area of about 1-m diameter represents the conditions across the site, including both sunlit and shaded (under grass or shrub cover) soils. Tuzet et al. (1997) described a ‘‘shade effect’’ that could affect the partitioning of ET due to variability of soil surface water availability. In 1990 at WGEW, electrical resistance sensor (ERS) soil moisture probes were deployed at Lucky Hills under three bare and three shrub-covered surfaces at 5-cm depths (Hymer et al., 2000). Due in part to the low vegetation cover in this region, the difference between surface soil moisture between bare and shrub-covered surfaces was only 0.002 $cm^3\,cm^{-3}$ over the entire 1990 study period (DOY 229–280). Similarly, the 20 microlysimeter measurements of soil evaporation over the 60 m × 60 m area varied by only 4% of the mean, when means ranged from 3 to 0.3 mm day^{-1} at both sites. This apparent uniformity of soil moisture and evaporation support the use of the IRT to represent soil conditions within the footprint of the Bowen ratio ET_D measurement at WGEW.

Another assumption is the value of E_{Dmin}. In the application to WGEW, we were able to set E_{Dmin} = 0.34 mm day^{-1} based on microlysimeter measurements of E_D during the North American monsoon season at both sites. Most locations would not have such information and an estimate of E_{Dmin} should be made or an energy balance model used for the extreme condition when soil moisture is low and yet ET_{PD} is high. On a given day, a small error in estimation of E_{Dmin} would not have a great impact on estimation of T_D; however, when summed over the study period, the error is cumulative and could be substantial.

5.2. Empirical supplement

A basic assumption used to derive the supplement is that estimates of T_D are correlated with another on-site measurement. In this study, we found that T_D was correlated with θ_{15} and we could use the relation between T_D and θ_{15} to estimate T_D for days when we could not assume $E_D \approx E_{Dmin}$. In other locations, this relation may not hold. For example, Kurc and Small (2004, 2007) reported that ET_D was related to θ_5 at other grass- and shrub-dominated locations. The most successful empirical supplement will be related to the wetting fronts of the soil and rooting depths of the plants. Alternatively, it may also be possible to use a theoretical supplement that could be calibrated with some empirical information gained from days when E_D is known to approach E_{Dmin}.

The relation between T_D and θ_{15} has a non-linear trend (Fig. 5) that was explained by Laio et al. (2001) and supported by Kurc and Small (2004). The selection of the ceiling value of T_D at 2.5 mm day^{-1} was based on the available data for three study periods in 2004–2006, including estimates of T_D at Kendall in 2005 by Green (2006). The upper limit for Lucky Hills was estimated to be near 3.0 mm day^{-1} using sap-flow gauges on selected shrubs in 2003 (Scott et al., 2006a,b). The uncertainty of that upper limit could contribute error to estimates summed over a prolonged study period and will likely have to be reassessed for application at other sites.

Like the Δt-based approach, the supplement is based on the assumption that the *in situ* measurement (in our study, θ_{15}) represents conditions within the footprint of the Bowen-ratio ET_D measurement. For the Kendall, Lucky Hills and other sites, Thoma et al. (2008) reported that the *in situ* measurements of θ_5 compared well with measurements made with a portable soil moisture sensor over the surrounding area of 35 m × 35 m with an average RMSE of 0.02 $m^3 m^{-3}$. Further, they reported that the measurements over 100 m × 100 m plots centered over three 35 m × 35 m plots differed by less than 2% when intra site θ_5 varied between 0.06 and 0.15 $m^3 m^{-3}$.

5.3. Operational application of the partitioning approach

The application of this approach in an operational mode over a multi-year period has some limitations. Since it is defined as a partitioning approach, one must assume that the vegetation is transpiring at a rate dependent upon soil water availability. Emmerich and Verdugo (2008a) reported that for the desert-adapted vegetation at Kendall and Lucky Hills, there is a precipitation threshold amount needed for a plant CO_2 uptake response and this threshold was different for spring and summer. They reported a lag between a precipitation event and the plant CO_2 uptake response following prolonged dry periods. This would conflict with the basic assumption of the supplement that T_D is linearly related to θ_{15} (Fig. 5). For this study, we limited analysis to a time when plants were already actively transpiring in the summer growing season and immediately responsive to changes in soil moisture.

For multi-year analysis of T_S/ET_S using this approach, the variation in site vegetation cover (or LAI) must be considered. For this study, the total vegetation cover at Kendall and Lucky Hills remained relatively unchanged over the 3-year period from 2004 to 2006 (King et al., 2008). This allowed us to compare T_S/ET_S for a variety of precipitation patterns for a given vegetation cover at each site. Further, the accuracy of the results depends on the accuracy of the estimate of ET using the Bowen ratio or eddy covariance instrumentation and theory (Twine et al., 2000). These are not limitations of this approach, but rather, general issues of concern with partitioning studies over multi-year time periods.

6. Summary and conclusions

A main contribution of this work was a new approach to partition on-site measurements of ET_D into daily E and T in a semiarid watershed based on the low-cost addition of an IRT to existing eddy covariance and Bowen ratio stations. We showed that reasonable estimates of T_D were obtained for clear-sky days when $E_D \approx E_{Dmin}$ based on the magnitude of the detrended Δt ($\Delta t'$). This finding is the most valuable and reliable aspect of this approach. The approach is based on instrumentation that can be maintained in place continuously for years with no more expertise and effort than is already required for deployment of energy flux stations.

An empirical supplement was added to compute T_D on days when the Δt-based approach is not applicable (e.g. cloudy days or when $E_D \neq E_{Dmin}$). The supplement used the relation between T_D and θ_{15} at the grass- and shrub-dominated sites to predict T_D. Then, the residual $ET_D - T_D$ difference was attributed to E_D, resulting in estimates of T_D and E_D over the entire study period. This approach provided reasonable results but tended to overestimate T_D. Alternatively, it may also be possible to use a theoretical supplement that could be calibrated with some empirical information gained from days when E_D is known to be near E_{Dmin}.

Using this new approach, we were able to study the sum of values over the study period (T_S/ET_S) for 3 years at one grass- and one shrub-dominated site at WGEW and address commonly held hypotheses. Basically, we found that T_S/ET_S was different at grass- and shrub-dominated sites and relatively insensitive to precipitation patterns. However, these conclusions must be taken in context. This study was limited to a study period (DOY 215–285) within the North American monsoon season for 3 years of near- to below-normal precipitation. Furthermore, the results are constrained by the fact that we are not studying shrub encroachment, but rather the differences between grass- and shrub-dominated sites at WGEW. Further studies of the impact of soil type on transpiration at grass- and shrub-dominated sites and the link between E_S and vegetation dynamics need to be undertaken to extrapolate the conclusions from the Kendall and Lucky Hills sites to provide a better understanding of shrub encroachment in grasslands.

All conclusions about the ecosystem dynamics of T_S/ET_S are premised on the accuracy and precision of this new approach for partitioning ET. Preliminary analysis suggests that the accuracy and precision of T_S are on the order of about 5% of T_S. Future work should focus on validation of the approach at WGEW and other locations. The approach will have to be refined for application to sites or seasons characterized by energy limiting conditions rather than water limitations. The significant change in the regional water cycle associated with vegetation transitions is an international concern in arid and semiarid regions, including Australia, Israel and the Middle East, and much of the continent of Africa (e.g., Eamus et al., 2006; Wood et al., 2007). With increased confidence, this approach could be used to better understand the potential hydrologic impact of native plant encroachment and exotic plant invasion at the landscape scale.

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