Evaluating the complementary relationship for estimating evapotranspiration from arid shrublands

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[1] Given increasing demands on finite water supplies, accurate estimates of evapotranspiration (LE) from arid shrublands of the Southwestern United States are needed to develop or refine basin water budgets. In this work, a novel approach to estimating the equilibrium (or wet environment) surface temperature (T_e) and LE from regionally extensive phreatophyte shrublands is tested using complementary theory and micrometeorological data collected from five eddy correlation stations located in eastern Nevada. A symmetric complementary relationship between the potential LE (LE_p) and actual LE is extremely attractive because it is based on general feedback mechanisms where detailed knowledge of the complex processes and interactions between soil, vegetation, and the near-surface boundary layer can be avoided. Analysis of computed LE_p and eddy correlation–derived LE indicates that there is unequivocal evidence of a complementary relationship between LE_p and LE, where the measured and normalized complementary relationship is symmetric when T_e is utilized to compute the wet environment LE (LE_w). Application of a modified Brutsaert and Stricker advection-aridity (AA) model, where T_e is utilized to compute LE_w as opposed to the measured air temperature, indicates an improvement in prediction accuracy over the standard Brutsaert and Stricker AA model. Monthly and annual predictions of LE using the modified AA model are within the uncertainty of the measurement accuracy, making the application of this approach potentially useful for estimating regional LE in arid shrubland environments. Our observational evidence supports the idea of a symmetric complementary relationship yielding an approach with standard parameters, making it simple to apply with satisfactory accuracy. To our knowledge, this work presents the first application and evaluation of the complementary relationship in phreatophyte shrublands while utilizing the T_e with comparisons to actual LE via flux measurements.

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1. Introduction

[2] Given increasing demands on finite water supplies in arid environments, the need for accurate estimates of sustainable groundwater resources is greater than ever. Many Great Basin and greater Southwestern United States drainage areas are considered hydrologically closed, where the entire groundwater recharge volume is consumed by evaporation and evapotranspiration along mountain front and valley floor areas. Because phreatophyte shrubs utilize shallow groundwater for transpiration, phreatophyte evapotranspiration is larger than direct precipitation. For example, in eastern Nevada it has been found that evapotranspiration from phreatophyte shrubs can range from 106% to 162% of the measured direct precipitation [*Moreo et al.*, 2007; *Welch et al.*, 2007].

[3] The amount of groundwater recharge that occurs in a given hydrographic basin is difficult to estimate accurately and is therefore commonly quantified by estimating the groundwater discharge for individual basins or entire flow systems if groundwater flows from one basin to another. Quantifying evaporation and evapotranspiration in the Great Basin has long been a major focus for developing and refining basin water budgets [*Maxey and Eakin*, 1949; *Robinson*, 1958; *Eakin*, 1966]. As such, many phreatophyte shrub evapotranspiration rates have recently been reassessed in the Great Basin region using micrometeorological, energy balance, and remote sensing techniques [*Malek et al.*, 1990; *Nichols*, 1994; *Tyler et al.*, 1997; *Nichols*, 2000; *Steinwand et al.*, 2006; *Moreo et al.*, 2007; *Allander et al.*, 2009]. More basic approaches have also been employed to estimate evapotranspiration (LE) from shrublands, such as multiplying the potential LE (LE_p) by the ratio of LE to LE_p, where micrometeorological, energy balance, and soil water

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balance methods are used to quantify this fraction [*Granger and Gray*, 1989; *Steinwand et al.*, 2001]. The LE in arid shrub environments is highly correlated to the amount of direct precipitation. Therefore, fractions of LE/LE_p covary in time and space with precipitation, making the application of LE/LE_p fractions for different time periods or areas of interest difficult and likely inaccurate without accounting for relative precipitation and soil moisture differences. Applying one- or two-source physically based models to estimate LE that consider energy transport from the soil and canopy [*Shuttleworth and Wallace*, 1985; *Kustas*, 1990; *Shuttleworth and Gurney*, 1990] is equally difficult to apply in arid shrub environments with confidence because of uncertainties in parameters relating complex aerodynamic, canopy, and soil resistances to sensible and latent heat fluxes [*Nichols*, 1992; *Stannard*, 1993].

[4] An approach based on general feedback mechanisms is attractive because it allows us to avoid extremely detailed knowledge of the complex processes and interactions between soil, vegetation, and the near-surface boundary layer. For this reason, methods that employ the complementary relationship (CR) of evapotranspiration have become popular, as they rely on feedbacks between LE and LE_p. The CR is related to water availability and near-surface atmospheric feedbacks with the land surface. Simply stated, when there is ample water available, LE increases and approaches the LE_p. When water is limited and the available energy is fairly constant in space, energy that would have been used for evapotranspiration is now used in the production of sensible heat flux and the vapor pressure deficit increases because of the lack of LE, thus elevating LE_p. *Bouchet* [1963] first hypothesized that there are complementary feedbacks between LE and LE_p, and related these fluxes to the available energy-limited wet environment LE, termed equilibrium evapotranspiration (LE_w). The equilibrium, or wet environment evapotranspiration rate, LE_w, is the LE_p of a wet surface having an area large enough to influence the atmospheric variables at a regional scale so that $LE_w \leq LE_p$. The complementary relationship can be expressed as

$$(1 + b)\mathrm{LE}_w = \mathrm{LE}_p + b\mathrm{LE}, \tag{1}$$

where b is a proportionality constant of unity if the relationship is symmetric. A symmetric CR implies that a unit increase in LE will result in a unit decrease in LE_p, and when the surface is saturated, LE = LE_w = LE_p (Figure 1). *Morton* [1969] and *Brutsaert and Stricker* [1979] further developed the idea and proposed a quantitative approach for estimating LE_p, LE_w, and LE on the basis of a symmetric CR and combination approach for estimating LE_p.

[5] The CR has been the subject of much debate regarding (1) whether the CR has physical basis and is actually complementary [*LeDrew*, 1979; *Lhomme and Guilioni*, 2006; *Szilagyi and Jozsa*, 2008; *Pettijohn and Salvucci*, 2009], (2) the cause of decreasing worldwide panevaporation during a period when air temperatures are increasing [*Brutsaert and Parlange*, 1998; *Roderick and Farquhar*, 2002; *Hobbins et al.*, 2004; *Ramírez et al.*, 2005; *Brutsaert*, 2006], and (3) recent findings that the CR is asymmetric for certain conditions [*Pettijohn and Salvucci*, 2006; *Kahler and Brutsaert*, 2006; *Szilagyi*, 2007; *Pettijohn and Salvucci*, 2009]. Despite some skepticism on its heuristic nature, the CR has been extensively applied to estimate regional-scale LE and has been tested against energy and large-scale water balance estimates of LE [*Morton*, 1983; *Brutsaert and Stricker*, 1979; *Hobbins et al.*, 2001; *Ozdogan and Salvucci*, 2004; *Kahler and Brutsaert*, 2006; *Yang et al.*, 2006; *Szilagyi and Jozsa*, 2008]. Results from these studies have all supported a realistic physical basis of the CR. Recent research has focused on various assumptions of model formulations, such as considering the stomatal conductance in the formulation of LE_p [*Pettijohn and Salvucci*, 2006], considering the wet environment temperature when estimating LE_w [*Szilagyi et al.*, 2009], and considering two-dimensional analytical and numerical modeling of the dry-wet interface [*Pettijohn and Salvucci*, 2009; *Szilagyi and Jozsa*, 2009a, 2009b].

2. Objectives

[6] This research utilizes a combined total of 10 years of micrometeorological data collected from five eddy correlation stations to test whether a CR exists in arid shrubland environments. This work also evaluates the performance of various CR formulations to predict LE from these environments. Assumptions in various formulations of the CR are explored by evaluating the shape and prediction accuracy of the CR. By considering the wet environment surface temperature in computing the LE_w, we demonstrate that the CR is indeed symmetric and prediction accuracy is improved over formulations that do not consider the wet environment surface temperature. Our observational evidence supports the idea of a symmetric CR yielding an approach with standard parameters, making it simple to apply with satisfactory accuracy. To our knowledge, this work presents the first application and evaluation of the CR utilizing the wet environment or equilibrium surface temperature with comparisons to actual LE via flux measurements.

3. Study Sites and Meteorological Data

[7] Study sites are located in eastern Nevada, within the Great Salt Lake and Colorado regional flow systems (Figure 2). The climate of the study sites is arid to semiarid, where the mean annual precipitation ranges from 150 to 250 mm with approximately 40% of the precipitation occurring in the winter months. The monthly average extreme temperatures range from 30°C in July to −10°C in December. The vegetation surrounding the study sites consists of spatially extensive and fairly homogeneous phreatophyte shrub species dominated by greasewood (*Sarcobatus vermiculatus*) with smaller amounts of rabbitbrush (*Chrysothamnus nauseous*), salt grass (*Distichlis spicata*), and sagebrush (*Artemisia tridentata*) (Figure 3), where the depth to groundwater ranges from 2 to 10 m below land surface [*Moreo et al.*, 2007]. Micrometeorological stations at the study sites were operated and maintained by the U.S. Geological Survey as part of the Basin and Range Carbonate-Rock Aquifer System Study [*Moreo et al.*, 2007; *Welch et al.*, 2007]. *Moreo et al.* [2007] computed LE at each site using the eddy correlation approach, where the average energy balance closure error for all sites averaged 10%. Daily average meteorological measurements of net radiation, ground heat flux, air temperature, vapor pressure, and wind speed are used in this study to estimate LE_p and LE_w, and the eddy correlation–derived LE is compared to CR-predicted LE. Energy balance closure corrections to LE and or sensible heat (H) were not performed

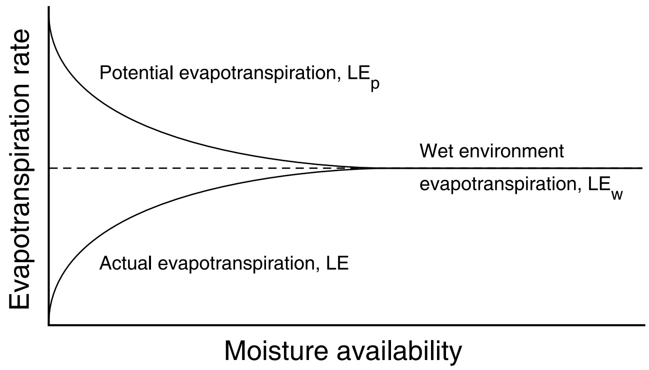

Figure 1. Conceptual representation of the complementary relationship of regional evapotranspiration. The farther to the right of the *x* axis, the wetter the regional environment, where LE increases and LE_p decreases.

because of relatively good closure in the original data (~10%) and uncertainties related to the measured available energy and closure correction procedures [*Twine et al.*, 2000; *Foken*, 2008]. For specifics regarding data processing and micrometeorological instrumentation at the study sites, refer to *Moreo et al.* [2007].

4. Advection-Aridity Approach

[8] The advection-aridity (AA) model proposed by *Brutsaert and Stricker* [1979] is based on a symmetric CR where *b* is unity and (1) becomes

$$LE = 2LE_w - LE_p. \quad (2)$$

The potential evapotranspiration, LE_p, expressed in terms of water depth equivalent of mm d^{-1}, is estimated by applying the combination approach by *Penman* [1948] to compute the potential evapotranspiration,

$$LEp = \frac{\Delta Q_n + \gamma E_a}{(\Delta + \gamma)}, \quad (3)$$

where Δ is the slope of the saturation vapor pressure curve at air temperature, γ is the psychrometric constant, and Q_n is the available energy at the surface expressed in terms of water depth equivalent of mm d^{-1}. E_a (mm d^{-1}) represents the drying power of the air and is expressed here using

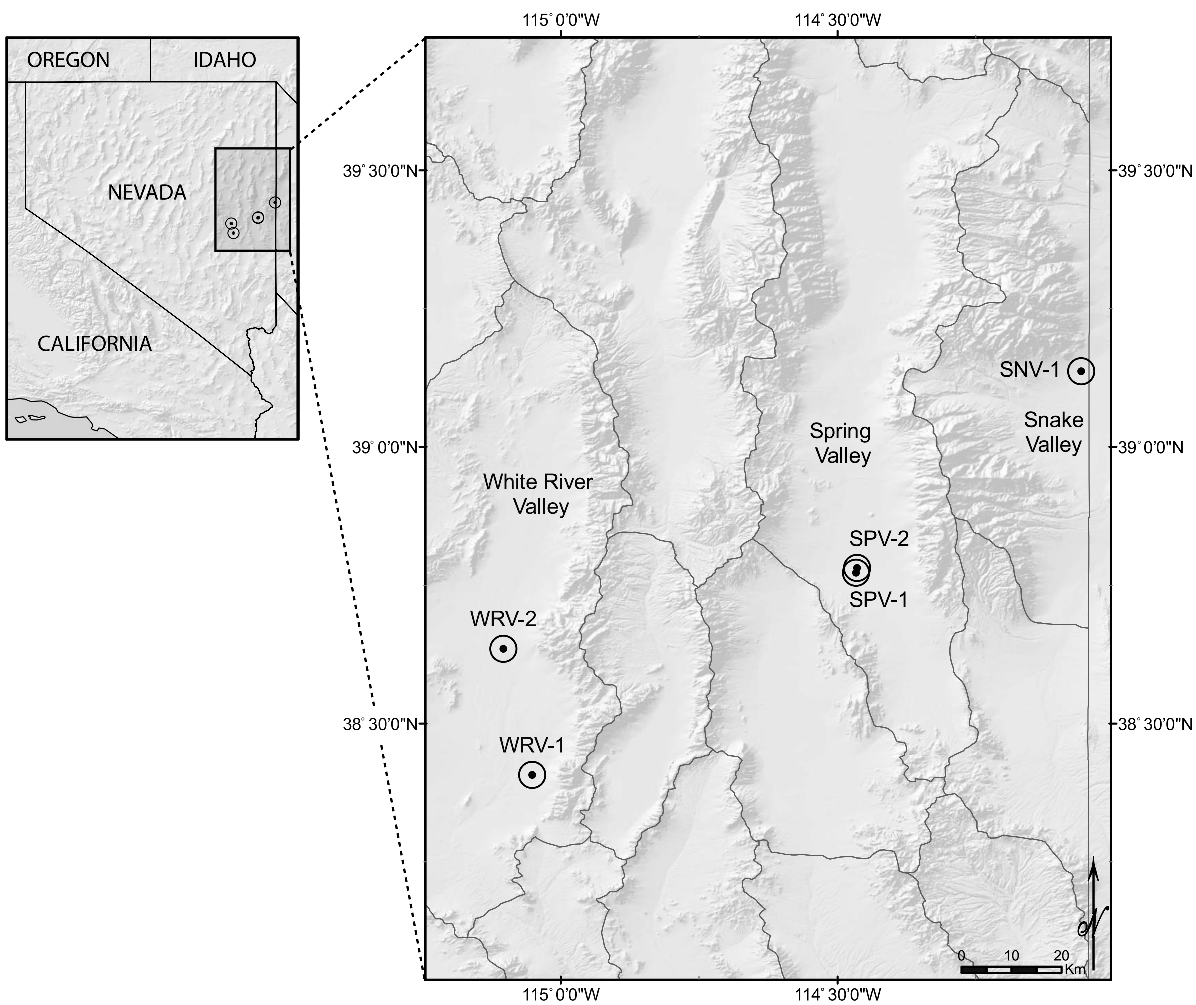

Figure 2. Site map of Nevada, White River Valley, Spring Valley, and Snake Valley eddy correlation stations operated and maintained by the U.S. Geological Survey [*Moreo et al.*, 2007] and used in this research for evaluating the CR at each site and collectively.

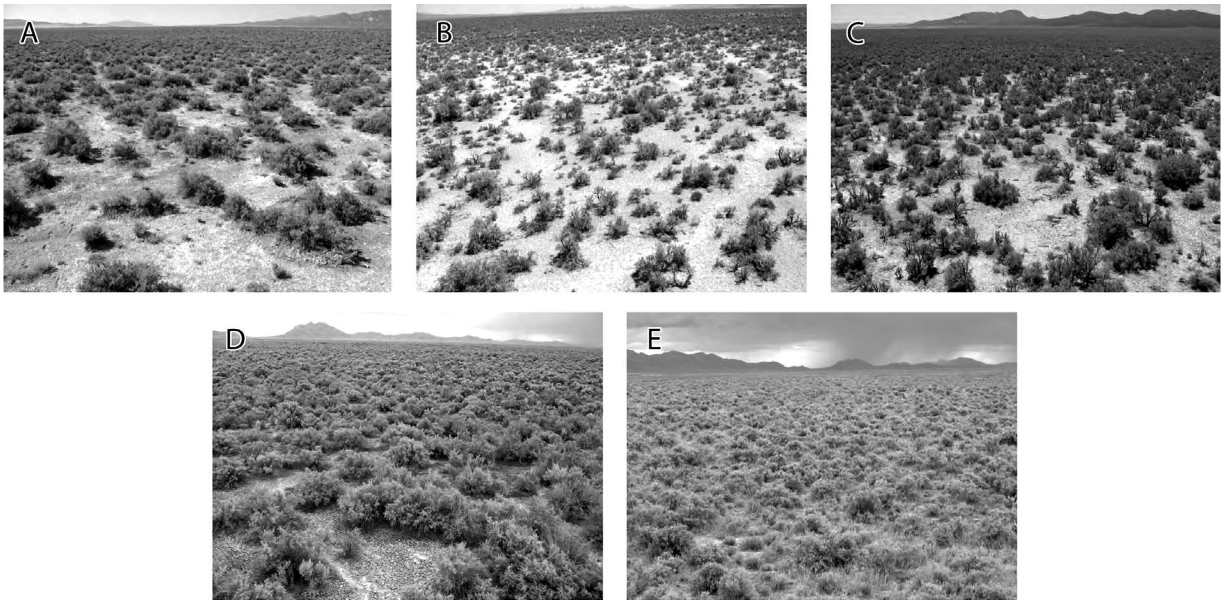

Figure 3. Site photos and primary vegetation type: (a) SNV-1 moderately dense greasewood, (b) SPV-1 sparse greasewood and rabbitbrush, (c) SPV-2 moderately dense grease wood and rabbitbrush, (d) WRV-1 dense greasewood, and (e) WRV-2 moderately dense greasewood. Photos modified from *Moreo et al.* [2007] (photos taken by Michael T. Moreo).

Penman's original Rome wind function for a wet vegetated or free water surface [*Brutsaert*, 1982] as

$$E_a = 0.26(1 + 0.54U)(e_s - e_a), \tag{4}$$

where e_s and e_a are the saturation and actual vapor pressures (hPa) and U is the measured wind speed (m s^{-1}) at a 2 m reference level. The Priestley-Taylor equation [*Priestley and Taylor*, 1972] is used to estimate the wet environment evapotranspiration at a length-scale greater than about 1 km as

$$\mathrm{LE}_w = \frac{\alpha \Delta Q_n}{(\Delta + \gamma)}, \tag{5}$$

where α is the well-known Priestley-Taylor coefficient. Commonly, α is used as a calibration coefficient; however, here α is fixed to the Priestley-Taylor original value of 1.26 to reduce the degrees of freedom.

4.1. The Modified Advection-Aridity Approach

[9] *Szilagyi and Jozsa* [2008] argue that Δ in (5) should be evaluated at the wet environment air temperature as opposed to the available (drying environment) air temperature since LE_w is intended to represent the wet environment LE. The wet environment air temperature is generally unknown under water-limited conditions but can be approximated by the wet environment surface temperature, T_e, because in wet environments the temperature gradient of the air is relatively small. T_e can be estimated iteratively by employing the Bowen ratio, B_o for a hypothetical small wet surface surrounded by water-limiting conditions so that ambient air temperature can be used:

$$B_o = \frac{H}{\mathrm{LE}_p} = \frac{Q_n - \mathrm{LE}_p}{\mathrm{LE}_p} = \gamma \frac{T_s - T_a}{e_s - e_a} \approx \gamma \frac{T_e - T_a}{e_s(T_e) - e_a}, \tag{6}$$

where H is the sensible heat; T_s and T_a are wet surface and measured air temperature, respectively; and $e_s(T_e)$ is the saturated vapor pressure taken at the wet environment surface temperature. By applying (3) with the measured Q_n, T_a, and e_a to estimate LE_p, all terms are known except for T_e and $e_s(T_e)$ and therefore can be solved iteratively. For T_e to be less than T_a, H is required to be negative, implying advection of energy over the hypothetical wet area. Equation (6) assumes that (1) the measured Q_n is spatially and temporally constant for each time step (daily in this case), which is valid given the large homogenous fetch at the sites; (2) the extent of the wet surface is small (making the Penman equation applicable with use of ambient weather data); and therefore, (3) measured air temperature and humidity over the surface are just minimally affected by the wet surface and can be estimated by the measured values under water-limited conditions. The key in the application of (6) is the realization that under a constant Q_n, required for the CR, the surface temperature of a small wet area would stay constant as the environment dries around it as shown by *Pettijohn and Salvucci* [2009] and *Szilagyi and Jozsa* [2009a, 2009b]. The modified AA model proposed by *Szilagyi and Jozsa* [2008] is identical to the original [*Brutsaert and Stricker*, 1979] except for using the iteratively solved T_e in computing the wet environment *LE* in (5), i.e.,

$$\mathrm{LE} = 2\mathrm{LE}_w(T_e) - \mathrm{LE}_p. \tag{7}$$

In arid environments, differences in computed Δ and LE_w using T_e versus T_a can be significant. While *Szilagyi et al.* [2009] successfully tested the modified AA model with water balance closure data from watersheds across the conterminous United States, (7) has not been validated with measured LE.

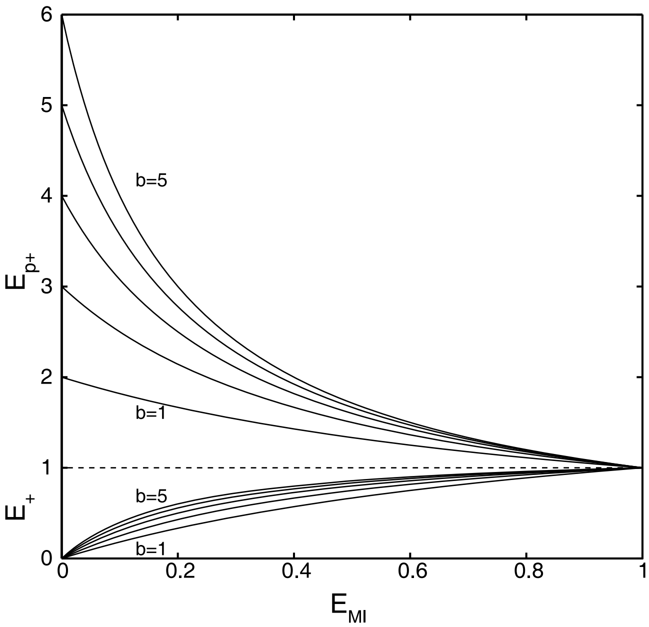

Figure 4. Theoretical normalized CR illustrating how different proportionality values of b impact equation (1). A b value of unity results in a symmetric CR. $E_+ = \mathrm{LE}/\mathrm{LE}_w$ and is considered the normalized LE, and $E_{p+} = \mathrm{LE}_p/LE_w$ is considered the normalized LE_p. E_{MI} is a normalized evaporative moisture index, where $E_{\mathrm{MI}} = \mathrm{LE}/\mathrm{LE}_p$. The dashed line is the equilibrium wet environment evaporation LE_w.

4.2. The Normalized Complementary Relationship

[10] Normalization procedures are attractive because they allow results or formulae to be expressed in a dimensionless form normalized by minimum or maximum values, where the minimum or maximum can change depending on location or environment. *Kahler and Brutsaert* [2006] normalized the CR by scaling LE and LE_p by LE_w, where $E_+ = \mathrm{LE}/\mathrm{LE}_w$ and $E_{p+} = \mathrm{LE}_p/\mathrm{LE}_w$. They formulate (1) as functions of the dimensionless variable (termed evaporative moisture index) $E_{\mathrm{MI}} = \mathrm{LE}/\mathrm{LE}_p$, where

$$E_+ = \frac{(1+b)E_{\mathrm{MI}}}{1+bE_{\mathrm{MI}}} \tag{8}$$

$$E_{p+} = \frac{1+b}{1+bE_{\mathrm{MI}}}. \tag{9}$$

Figure 4 illustrates the normalized CR where the scaled LE and LE_p are functions of E_{MI} (i.e., (8) and (9)). As the environment experiences wet surface conditions, E_{MI} increases to unity, where LE and LE_p approach LE_w. Conversely, as E_{MI} approaches zero, the environment experiences drying conditions where LE and LE_p diverge from LE_w. As shown in Figure 4, b is a proportionality constant controlling the shape of the CR. In their application of the normalized CR to pan data and Bowen ratio flux measurements, *Kahler and Brutsaert* [2006] recommended a b value of 5. The value of b has been described for an evaporation pan as a measure of the energy transfer between the pan and the surrounding environment [*Brutsaert*, 2006]. When applied to pan data, the CR is asymmetric ($b > 1$) because of the fact that the pan is exposed to more energy than the surrounding environment via radiation, conduction, and advection, and has increased mass transfer because of its small size [*Kahler and Brutsaert*, 2006; *Brutsaert*, 1982].

[11] In this work, it is shown that by evaluating (5) at T_e and estimating LE_p via the Penman equation, b becomes unity, yielding a symmetric CR. Whether (5) is to be evaluated at T_a or T_e becomes an issue only in arid environments where the $T_a - T_e$ difference can be large [*Szilagyi et al.*, 2009]. Figure 5 illustrates average monthly T_a and computed T_e for each site, where it is evident that T_e differs significantly from T_a as T_a becomes large. Making the CR symmetric through the estimation of T_e is advantageous because it eliminates the need to calibrate α and/or b. Note that the application of pan data in the CR for estimation of LE_p also requires additional meteorological data (air temperature and radiation) for estimating LE_w, and it requires a calibrated pan coefficient C_p, and/or b because of the pan's extreme sensitivity (b = 4 ~ 10) to changes in energy between the pan and the surrounding environment.

5. Data Preparation

[12] Meteorological data from five sites (Figure 2) were acquired from the U.S Geological Survey and were aggregated from 30 min to daily time steps. These data were used in (3), (4), and (5) to compute LE_p and LE_w, respectively. There has been debate about the appropriate temporal resolution for which to apply the CR. *Brutsaert and Stricker* [1979] and *Morton* [1983] recommend that the minimum temporal resolution for which the CR should be applied is 3–5 days because of passing weather fronts disrupting the dynamic equilibrium between the boundary layer and the environment. *Szilagyi and Jozsa* [2008] concluded that application of the AA model at daily or monthly time steps did not affect the predicted monthly accumulated LE values

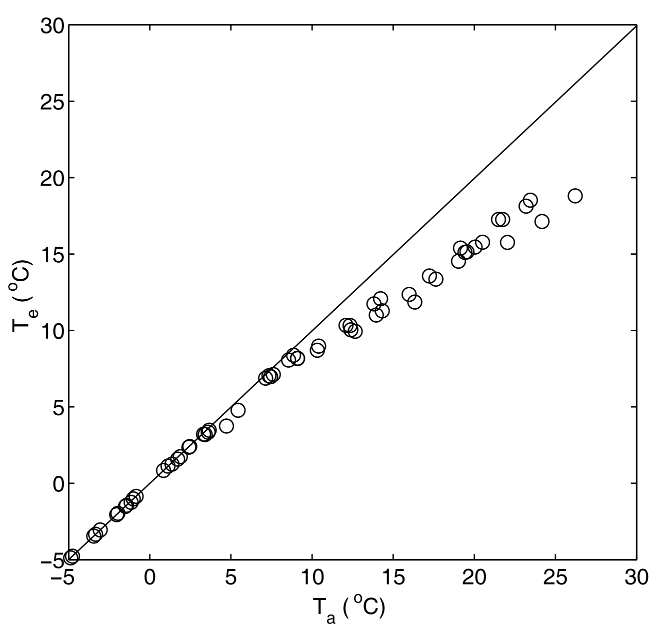

Figure 5. Average monthly measured air temperature, T_a, and computed T_e from inversion of (6). During warm season months, differences between T_a and T_e become large. T_e was assumed to equal T_a if T_e was greater than T_a.

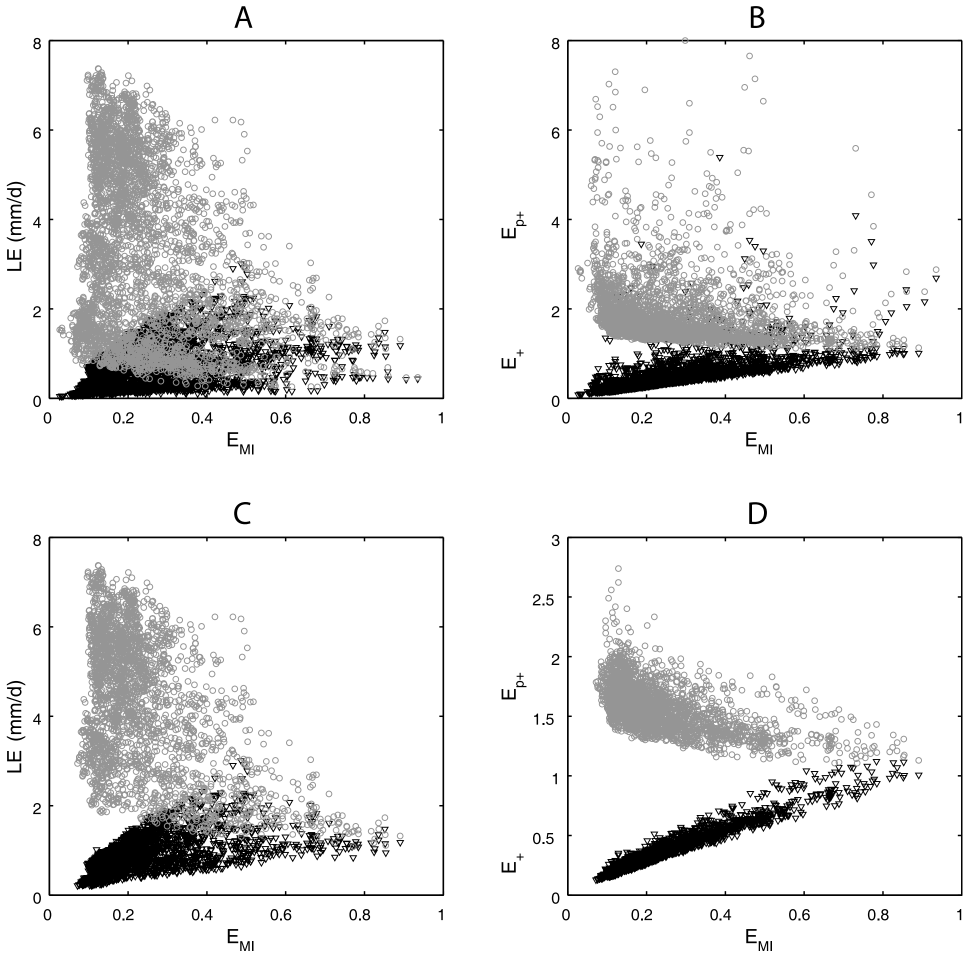

Figure 6. The nonnormalized and normalized CR for different time periods. (a) Daily LE_p (gray circles) and daily LE (black inverted triangles) for all data from all five sites. (b) Daily normalized LE_p as E_{p+} (gray circles) and daily normalized LE as E_+ (black inverted triangles) for all data from all five sites. (c) Daily LE_p and daily LE for March–November data from all five sites. (d) Daily normalized LE_p as E_{p+} and daily normalized LE as E_+ for March–November data from all five sites, where it is evident that a clear and concise CR exists once the data are filtered and normalized.

significantly. *Kahler and Brutsaert* [2006] applied their optimized CR at a daily time step using pan data but eliminated days where advection of moist air from a nearby reservoir was suspected. To provide continuity and increase the number of observations and predictions to analyze, we chose to average daily LE_p, LE_w, $LE_w(T_e)$, and measured LE using a 7 day moving window centered on the fourth day for computing E_+, E_{p+}, and the predicted LE.

6. Searching for a Complementary Relationship

[13] Potential complementary feedbacks that occur between the atmosphere and the surrounding environment are explored by analyzing the LE_p and measured LE as a function of the E_{MI} for all five sites. Daily 7 day average LE_p and LE versus E_{MI} spanning from approximately 15 August 2005 to 31 August 2007 for all five sites are plotted together, making a total of 3718 days (Figure 6a). Figure 6a illustrates that there is indeed complementary behavior between LE_p and LE; however, data points are quite scattered. As stated by *Kahler and Brutsaert* [2006], a truly universal relationship requires that the formulation be dimensionless, which in this case is accomplished by scaling the LE_p and LE by the LE_w. Scaled LE_p and LE as a function of the E_{MI} is shown in Figure 6b, where it is clearly evident that a complementary relationship between LE_p and LE exists. Still, there is a large degree of scatter and asymmetric behavior in E_{p+}. This scatter occurs during winter months when the LE_w is very small (0.05–0.1 mm d^{-1}) and LE_p is relatively large (0.5–1 mm d^{-1}), which inflates the LE_p/LE_w ratio (i.e., E_{p+}), resulting in asymmetry in the CR. As discussed in section 5, weather fronts can decouple the dynamic equilibrium between the land surface and the atmosphere. Most weather fronts occur over the study area during winter periods, while spring and summer months are relatively calm and are often associated with fairly consistent high-pressure weather patterns. Because extremely small values of LE_w occur in the denominator of E_{p+} during winter periods, data from winter periods (December–February) were eliminated. Figure 6c illustrates a more coherent CR between LE and LE_p after removing data during winter periods. Roughly 90% of the total LE from the study sites occur during March–November, which for all practical purposes warrants the exclusion of winter periods for quantitative analysis of the

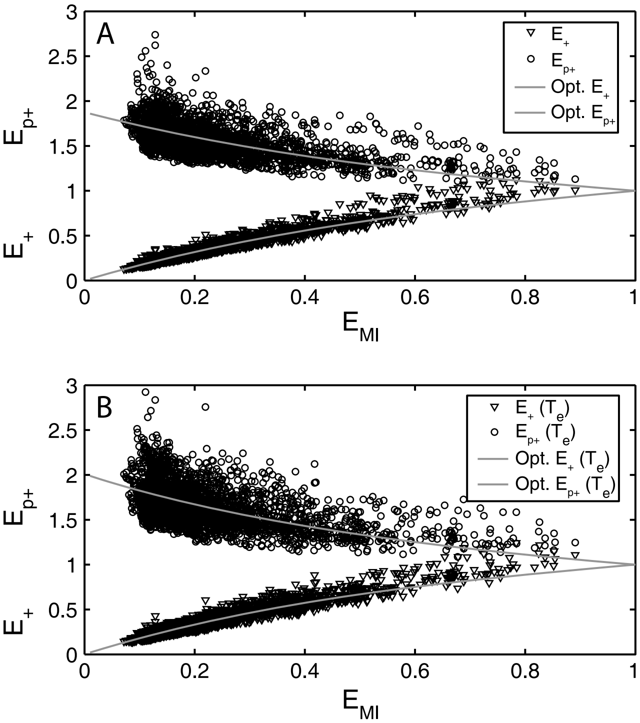

Figure 7. Measured normalized CR (black dots) and optimized theoretical CR curves using (a) the measured air temperature for estimating LE and (b) the iterated equilibrium temperature, T_e. The optimized proportionality constant, b, was found to equal 0.81 using T_a and was found to equal 1.00 using the iterated T_e.

shape of the CR and the calibration of b. Eliminating winter periods and normalizing LE and LE_p by LE_w results in unequivocal evidence that a CR exists in these environments (Figure 6d).

6.1. Asymmetric or Symmetric CR

[14] A symmetric CR is desired, not only to provide a consistent and unified theory that explains feedback processes between the land surface and the atmosphere, but also to provide consistency and simplicity for predicting LE. *Kahler and Brutsaert* [2006] show that the increase in available energy that an evaporation pan experiences causes the CR to become asymmetric. *Szilagyi* [2007] suggested that whenever advection of energy is present around the device (such as a class A pan) that estimates LE_p, the time rate of change between LE and LE_p is not a constant but is a function of the surface temperature. *Pettijohn and Salvucci* [2006] found that nonconvergence and asymmetry of the CR exists when canopy conductance is not considered in estimating LE_p. They recommend that the Penman-Monteith equation be employed with specified maximum conductance terms and stability correction to reduce over estimation of LE_p, and hence under estimation of LE.

[15] Rather than explore how variations in estimating LE_p affect the shape of the CR and prediction accuracy, as was done previously, we chose the *Penman* [1948] formulation of LE_p to explore the CR with estimated values of LE_w using the wet environment surface temperature. The shape of the CR is evaluated for spring, summer, and fall months for all five sites combined by optimizing the proportionality factor b in (8) and (9) simultaneously. In this analysis, the sum of square errors is minimized between 7 day average predicted and measured E_+ and E_{p+}. T_e is iteratively solved by employing (6), the Bowen ratio for a small wet surface using daily average measured air temperature and vapor pressure. Results indicate that $b = 0.810$ using LE_p and LE_w and $b = 1.008$ using LE_p and $LE_w(T_e)$ for all sites combined (Figure 7). Residuals of predicted E_+ and E_{p+} about the theoretical normalized CR curves were found to be correlated to net radiation, R_n, but not U, T_a, or e_a. Residuals became positive for higher R_n values and negative with larger variance for lower R_n values, indicating that errors were largest during spring and fall periods. Unlike the findings by *Pettijohn and Salvucci* [2006], when the standard forms of the Penman (with the Rome wind function) and Priestley-Taylor (α = 1.26) equations are employed to estimate LE_p and LE_w, a unit decrease in the LE_p results in more than a unit increase in LE. Of significant interest is the fact that b = 1.008, indicating a symmetric CR when the estimated wet environment surface temperature is utilized for computing LE_w. Because (5) is "expecting" a wet environment temperature, it seems that the use of T_e in (5) is more appropriate for applications of the CR in arid environments because of its reliance on the wet environment LE.

7. Application and Results

[16] Application and evaluation of the CR approach to estimate LE have been explored by several investigators at various temporal scales of annual [*Morton*, 1983; *Hobbins et al.*, 2001, 2004], monthly [*Morton*, 1983; *Xu and Singh*, 2005; *Szilagyi and Jozsa*, 2008; *Szilagyi et al.*, 2009], daily [*Brutsaert and Stricker*, 1979; *Crago and Brutsaert*, 1992; *Kahler and Brutsaert*, 2006; *Granger and Gray*, 1990; *Pettijohn and Salvucci*, 2006; *Crago and Crowley*, 2005; *Qualls and Gultekin*, 1997; *Granger and Gray*, 1989; *Szilagyi*, 2007], and 20 min periods [*Parlange*

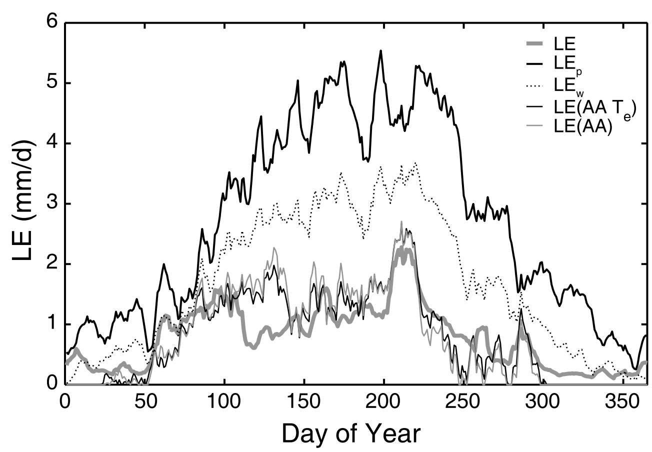

Figure 8. Time series of 2006 SPV-2 7 day moving average measured LE, LE_p, and LE_w; estimated AA LE using LE_w and LE(AA) (2); and estimated AA(T_e) LE using $LE_w(T_e)$ and LE(AA T_e) (7). LE_w is computed using the measured air temperature, whereas $LE_w(T_e)$ is computed using the iterated T_e.

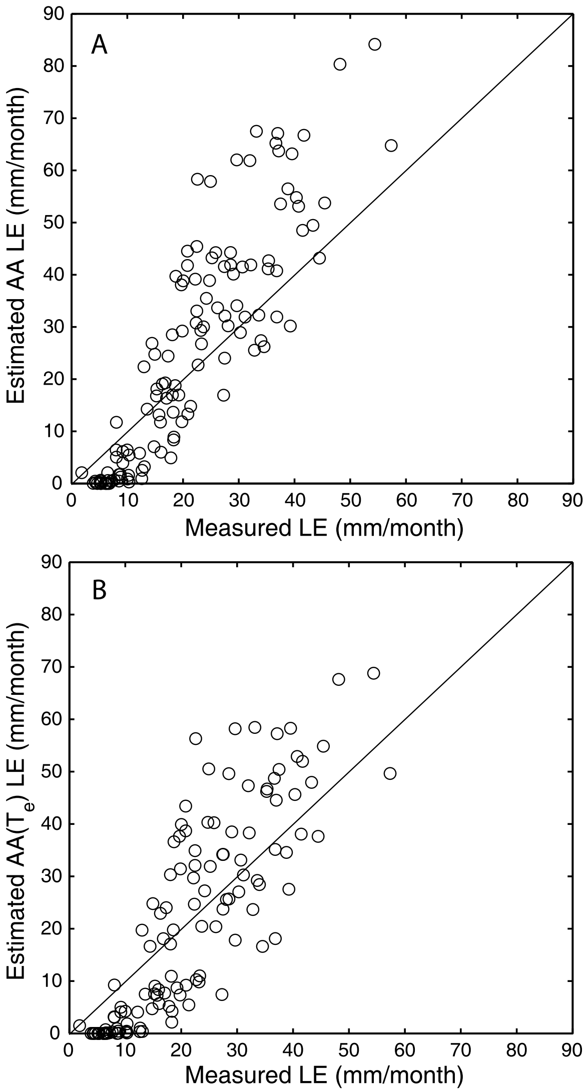

Figure 9. Measured versus estimated monthly LE using the (a) AA and (b) AA(T_e) models for all sites. The positive bias in Figure 9a is reduced when the equilibrium temperature T_e is used to compute LE_w in the AA model (Figure 9b).

and Katul, 1992]. In this research, application of 7 day moving average LE_p, LE_w, and $LE_w(T_e)$ are applied to (2) and (7) (i.e., AA and modified AA models, respectively), and the predicted LE is compared to the measured LE for all five sites. A time series comparison for site SPV-2 is shown in Figure 8, where both approaches predict rapid temporal changes in LE reasonably well considering that predictions are simply relying on symmetric feedbacks between LE and LE_p. However, there are periods where LE is overpredicted, mainly during spring months. In winter months the predicted LE is often negative as a result of the LE_p exceeding the LE_w by a factor of 2. These results are generally consistent among all sites.

[17] Underpredictions occur during winter months primarily because of dry, windy conditions as winter weather fronts pass through the study area. These conditions elevate the Penman equation while suppressing the Priestley-Taylor equation as a result of low available energy. For these reasons negative LE predictions were assumed to be zero. Figure 9 illustrates 1:1 plots of measured and predicted monthly LE using (2) and (7), where R^2 is 0.77 and 0.71 for the AA and AA(T_e) models, respectively. Although the LE computed with the AA(T_e) model has more scatter because of uncertainties in the iteratively computed wet environment temperature, the predicted LE is closer to the 1:1 line when compared to the AA model–predicted LE. The average monthly percent bias for all sites improves from 18% to 1% when LE is computed with the AA(T_e). The average monthly root-mean-square error for all sites is 11 and 13 mm for the AA(T_e) and AA models, respectively. Figure 10 illustrates the annual total LE for approximate water years (1 October to 30 September) for years 2006 and 2007, where it is evident that the inclusion of the wet environment temperature in the AA(T_e) model improves the annual total LE as compared to the AA model. Considering that

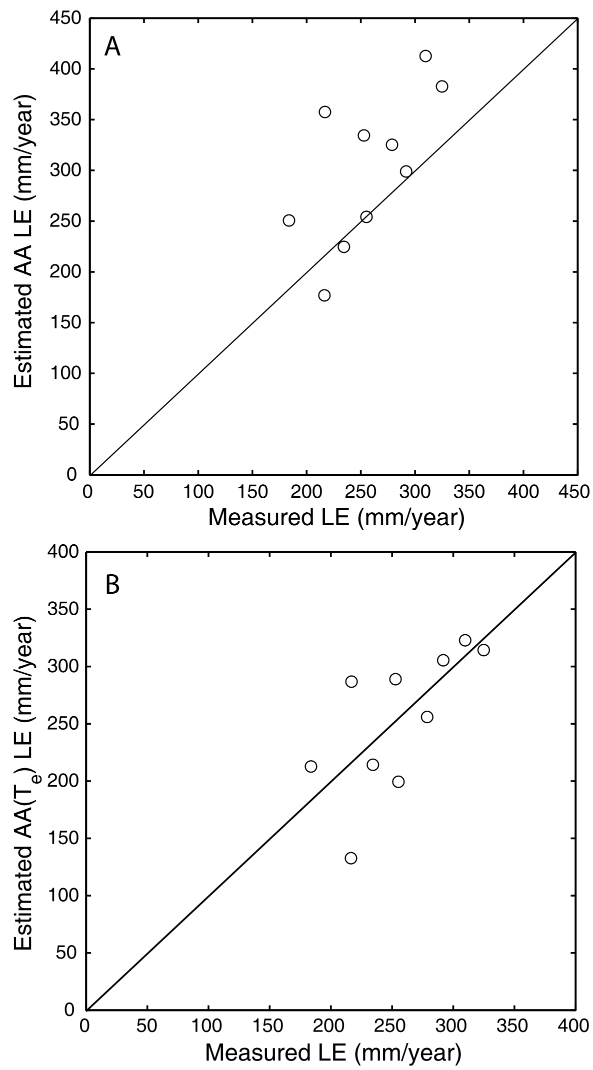

Figure 10. Measured versus estimated annual LE using the (a) AA and (b) AA(T_e) models for all sites. The positive bias in Figure 10a is reduced when the equilibrium temperature T_e is used to compute LE_w in the AA model (Figure 10b). Predictions are generally within the uncertainty of the measurements estimated to be 10% [*Moreo et al.*, 2007].

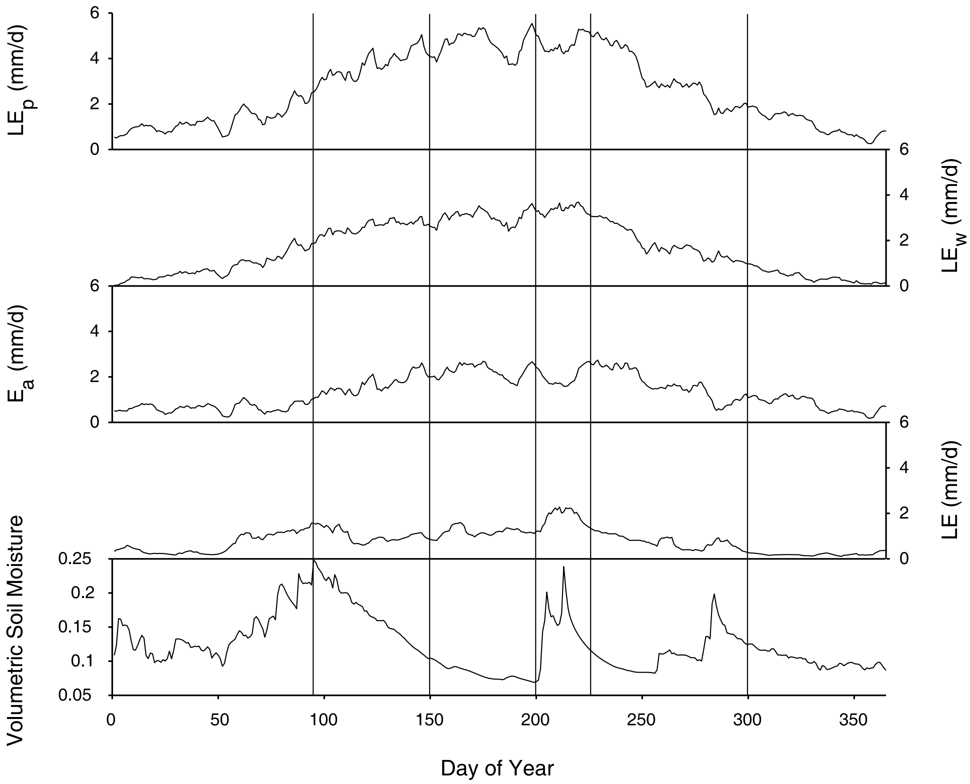

Figure 11. Comparison between SPV-2 2006 time series of LE_p, LE_w, drying power of the air, E_a, measured *LE*, and measured and volumetric soil moisture at 15 cm depth. Vertical solid lines indicate periods of interest where soil moisture changes are both positively and negatively correlated to LE, E_a, and LE_p. Note the clear response of E_a and hence LE_p to increased soil moisture and LE during days 200–225. This rapid response in E_a leads to fairly accurate prediction of LE during this period using AA models as shown in Figure 8.

the measurement error is approximately 10%, most of the predicted annual LE using the AA(T_e) model is generally within the measurement error.

8. Discussion

[18] Regional estimates of LE using the CR in arid environments require that E_a be sensitive to fairly rapid changes in surface moisture conditions. A time series of LE_p, LE_w, E_a, eddy correlation–derived LE, and available volumetric soil moisture measured at 15 cm depth at SPV-2 is illustrated in Figure 11, where it is evident that the LE_p, E_a, and LE are slightly positively correlated to changes in measured soil moisture in winter months because of non-water-limited conditions. As the LE_p and LE_w increased during spring months (days 90–150) because of increased Q_n and E_a, LE was fairly constant and insensitive to the large decrease in soil moisture. The fairly constant rate of LE during this period was likely due to the utilization of shallow groundwater, where the depth to water at SPV-2 was at an annual minimum of 2.1 m below land surface. During early summer (days 150–200) the soil moisture was continuing to decrease, while the LE was slightly increasing because of the general increase in Q_n and E_a and the utilization of shallow groundwater by phreatophyte taproots. During midsummer (days 200–225), soil moisture significantly increased because of summer monsoon rains, LE reached the annual maximum, and E_a and hence LE_p was significantly reduced because of the increase in LE, while LE_w remained fairly constant. Late summer and early fall months (days 225–300) experienced a general decrease in LE, which follows the general decrease in LE_w, E_a, and hence LE_p. During the later portion of this period there is a marked increase in soil moisture and slight increase in LE. These rapid land surface and lower boundary layer feedbacks are remarkably captured by the AA model in the prediction of LE as illustrated in Figure 8. Annual precipitation for these sites ranged from 150 to 250 mm yr^{-1} during the study period, where the annual measured and predicted LE ranged from 170 to 400 mm yr^{-1} (Figure 10), indicating consumption of shallow groundwater. The AA model is able to predict the utilization of shallow groundwater through the reduction in the measured vapor pressure deficit and LE_p. Both seasonal and annual results of this work serve as prime examples of the fundamental concept, fairly robust prediction accuracy, and merit of applying the CR in areas where LE and LE_p are correlated not only with precipitation [*Yang et al.*, 2006] but other sources such as shallow groundwater.

[19] Temporal variations in soil moisture have been used to scale LE_w or LE_p as a means to estimate LE from crops [*Davies and Allen*, 1973], forests [*Flint and Childs*, 1991; *Black*, 1979], and desert vegetation [*Garcia et al.*, 2009]. Rapid increases of LE commonly occur in arid and semiarid environments during spring and summer months because of precipitation events, but these events are not always captured in measured soil moisture at shallow depths. *Garcia*

et al. [2009] developed a function relating measured soil moisture to inversely calibrated α values and used these soil moisture–dependent α values to scale LE_w for prediction of bare soil LE in southern Nevada. *Garcia et al.* [2009] reported α values ranging from 0.25 to 1.4 for nearly constant volumetric soil moisture of 0.05 and α values ranging from 0.75 to 1.75 for nearly constant volumetric soil moisture of 0.20. *Black* [1979] also showed this large variability during dry periods, where LE/LE_w ranged from 0.2 to 0.9 for volumetric soil moistures of 0.12 to 0.15 because of precipitation events. Measured soil moisture has been used to develop various forms of the CR as well [*Crago and Brutsaert*, 1992; *Kahler and Brutsaert*, 2006; *Pettijohn and Salvucci*, 2006; *Yan and Shugart*, 2010]; however, similar to findings of *Kahler and Brutsaert* [2006], the use of soil moisture measurements at 15 cm depth to supplement E_{MI} in this research did not yield usable results for reasons related to high winter time soil moisture storage and low LE_w with relatively high corresponding LE_p from passing weather fronts, and rapidly varying soil moisture conditions in the summer time. These findings suggest that measured soil moisture at one depth does not properly characterize the surface soil moisture status, nor does it characterize the rapidly varying evaporative and transpirative conditions that exist in water-limited environments.

[20] The use of T_e in (5) to estimate the LE_w is of particular importance in arid environments because $T_a - T_e$ can be quite large, as shown in this work. E_a was originally calibrated with ambient measurements using the available T_a and humidity near the experiment site [*Penman*, 1948]. Had these measurements been taken directly over a large free water surface, the parameters of the wind function in (4) would be different. In contrast, (5) was calculated and α was optimized under actual regionally wet conditions [*Priestley and Taylor*, 1972]. This leads to the reason why in arid environments (3) does not require T_e and why (5) is more accurately estimated with T_e. Our findings demonstrate this argument by introducing T_e in (5) for estimating LE_w, where the CR becomes symmetric from a slightly asymmetric CR using T_a in (5) where the value $b = 0.876$. This finding is consistent with the idea that advection of energy at the study sites is negligible, otherwise the value of b would exceed unity, as *Kahler and Brutsaert* [2006] point out. This leads to a logical explanation of why b would have a value less than unity: the estimated LE_w was artificially inflated by using the measured T_a.

9. Conclusions

[21] Quantifying monthly and annual rates of LE in arid shrub environments is important for updating and developing groundwater budgets in the Southwestern United States. In this study, we demonstrate clear evidence of a CR between LE and LE_p in arid shrublands by utilizing eddy correlation data and commonly measured weather variables. We show that the CR is fairly robust for predicting rapid changes in LE, as well as total monthly and annual LE rates; however, winter predictions are underestimated. Furthermore, we show that by employing the wet environment temperature, T_e, for estimating LE_w, the optimized CR becomes symmetric, reduces the bias, and improves the accuracy of the total monthly and annual predicted LE when compared to eddy correlation–derived LE. The fact that CR is symmetric in arid shrub environments, where b equals unity and α is the original quantity of 1.26, leaves no calibration parameters to estimate LE and requires only commonly measured weather data and measured or predicted Q_n.

[22] Application of the CR can be used to study complex feedbacks between the land surface and near-surface boundary layer and to complement other studies such as remote sensing, vegetation phenology, and regional-scale hydrologic and atmospheric modeling. This paper summarizes the first application of the CR to estimate LE from phreatophyte shrubs and, it is hoped, will spur wider application to better understand and predict hydroclimatology in arid environments.

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