

## TRENDS IN REGIONAL EVAPOTRANSPIRATION ACROSS THE UNITED STATES UNDER THE COMPLEMENTARY RELATIONSHIP HYPOTHESIS

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**Abstract.** The hypothesis of a complementary relationship in regional evapotranspiration allows for estimation of actual evapotranspiration on a regional scale by simple, physically based models that take into account feedbacks in land surface-atmosphere dynamics. A regional, seasonal Advection-Aridity model is used to create a monthly time-series of actual evapotranspiration for a period of 27 years at a 5-km resolution over the conterminous United States.

This time-series allows trend analysis of actual evapotranspiration using the Mann-Kendall test on annual and seasonal bases and results are presented for the conterminous United States. Trends are analyzed within the context of the complementary relationship on a regional basis to establish that regional trends can be determined to originate in either the energy budget or the water budget, or both. Intra-annual trend results are compared with results from another study.

The regional-seasonal Advection-Aridity model is offered as a tool for studies of climate change and variability.

### 1. Introduction

In the study of the hydrologic effects of climate change and variability on the watershed-scale, it has long been acknowledged that not only is actual evapotranspiration ( $ET_a$ ) the component of the surface hydrological cycle that presents the greatest modeling challenge, but it is also that component which is the most sensitive to climate change and variability.

Other studies have concentrated on the directly observable components of the hydrological cycle; namely precipitation [Karl and Knight, 1998] and streamflow [Wahl, 1992; Chiew and McMahon, 1996; Lins, 1997; Lins and

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Slack, 1999]. Even those studies that take a more holistic hydrological approach—whether it be an examination of indices of climate change over the conterminous United States [Karl *et al.*, 1996], of long-term hydroclimatological trends over the conterminous United States [Lettenmaier *et al.*, 1994], or the effects of climate change on the hydrology of a smaller, regional-scale basin [Westmacott and Burn, 1997]—do not examine trends in  $ET_a$ . When evapotranspiration has been studied, it is in the context of potential evapotranspiration ( $ET_p$ ) at a point [Lockwood, 1994]—and that often derived from observations at evaporation pans [Eitzinger *et al.*, personal communication, 2001]—and not  $ET_a$  at a regional scale. While such studies are of interest in the context of trends in the standard agrometeorological variables, wherein  $ET_a$  is taken to be a function of  $ET_p$  determined by coefficients that are purported to describe local crop and soil moisture conditions, they neglect the complementarity of  $ET_a$  and  $ET_p$  evident at the regional scale. Thus, there is a need for analyses of trends in  $ET_a$ , whether they be due to climate change and variability or land-use change, on a seasonal temporal basis and regional scale useful to water managers looking to resolve issues of water supply and demand in increasingly variable hydroclimatic conditions.

Simply put, the primary reasons for this dearth of direct analysis of  $ET_a$  are the difficulties inherent in calculating  $ET_a$  and the complications of scale. In the study of evapotranspiration, the physics of energy and mass transfer at the land surface-atmosphere interface can be successfully modeled at small temporal and spatial scales [Katul and Parlange, 1992; Parlange and Katul, 1992a; Parlange and Katul, 1992b], yet our understanding of these processes at the larger—basinwide or regional—scale necessary for use by hydrologists, water managers, and climate and ecological modelers is limited. Consequently, our ability to estimate regional evapotranspiration is often constrained by models that treat  $ET_p$  as an independent climatic forcing process and derive  $ET_a$  estimates through non-physical, locally parameterized relationships thereof.

Thus,  $ET_a$  is generally modeled either as a simple, often empirical function of pan evaporation observed at nearby weather stations, or by ground-based models that tend to rely on gross assumptions regarding processes in the poorly understood soil, vegetative, and atmospheric systems, while ignoring the many feedback mechanisms and interactions that occur at the interfaces between the three [Morton, 1983].

Most soil systems in current watershed models invoke the assumption that moisture fluxes across the soil surface-atmosphere interface are assumed to be independent of temperature or gradients of temperature and vapor pressure.

Vegetation is generally assumed to act as a “passive wick” in the transfer of moisture from tension storages to the atmosphere, thereby ignoring counter-intuitive feedback loops at the vegetation-atmosphere interface that reduce the actual transpiration rate in response to increases in the vapor pressure deficit of the ambient air and corresponding increases in  $ET_p$  [Morton, 1983].

With regard to the atmospheric assumptions of the soil-plant-atmosphere system,  $ET_p$  is traditionally assumed to be independent of  $ET_a$  and is used to

reflect evaporation demand. It is often calculated by approximate solutions such as the Penman equation. However, it is well known that the passage of air masses over continents modifies them according to the proportion of net radiative energy taken up by  $ET_a$  and thus transferred as latent heat back to the atmosphere. Decreases in  $ET_a$  due to reductions in water availability lead to temperature increases and humidity decreases of the over-passing air, which lead to an increase in the drying power of the air.

These gross assumptions as to the nature of moisture dynamics in each of the components of the land surface-atmosphere interface, and of the interactions between them, point to the need for models of regional evapotranspiration that consider the evaporative system as an integrated whole, that bypass the poorly understood dynamics within each component, and that incur minimal data requirements as to the nature of the land surface. Complementary relationship models are such models.

In this paper, then, the authors seek to present a complementary relationship model as a tool for climate change studies and to present the initial findings of a study of trends in  $ET_a$  across the conterminous US. The  $ET_a$  data is drawn from one of the most common complementary relationship formulations: the Advection-Aridity approach first proposed by *Brutsaert and Stricker* [1979] and since reparameterized on a regional and seasonal basis by the authors [*Hobbins et al.*, 2001b]. This model combines the effects of regional advection on  $ET_p$  with the hypothesis of a complementary relationship between  $ET_p$  and  $ET_a$ .

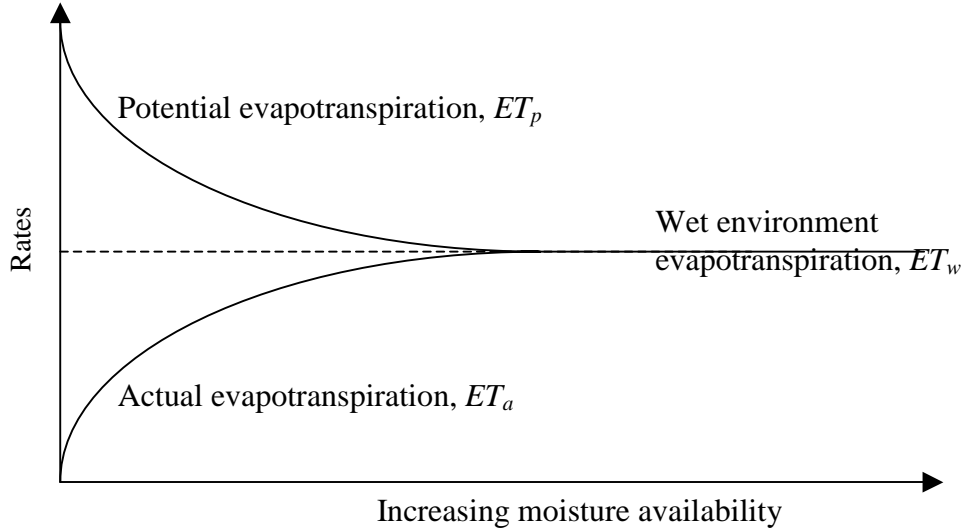
## 2. The Complementary Relationship

The theory of the complementary relationship in regional evapotranspiration, first proposed by *Bouchet* [1963], considers  $ET_p$  a function of the aforementioned feedback processes between limitations of water availability at the land surface and the evaporative power of the overlying atmosphere, while avoiding the difficulties inherent in coupling the microscale (surface) and macroscale (free atmosphere) phenomena in the soil-plant-atmosphere system. In modeling the complementary relationship,  $ET_a$  is estimated using only data that describe the conditions of the over-passing air and that are already measured for  $ET_p$ , and no locally optimized coefficients are necessary.

The hypothesis of a complementary relationship states that over areas of a regional scale and away from any sharp environmental discontinuities, there exists a complementary feedback mechanism between actual and potential evapotranspiration rates [*Bouchet*, 1963]. Energy at the surface that, due to limited water availability, is not taken up in the process of  $ET_a$  increases the temperature and humidity gradients of the over-passing air and leads to an increase in  $ET_p$  equal in magnitude to the decrease in  $ET_a$ . This relationship is described by equation (1):

$$ET_a + ET_p = k.ET_w. \quad (1)$$

In homogeneous regions where the assumptions of a well-mixed atmospheric boundary layer and negligible upwind edge-effects hold well, the value of  $k$  has generally been taken to equal two. Under conditions where  $ET_a$  equals  $ET_p$ , this rate is referred to as the wet environment evapotranspiration ( $ET_w$ ). Figure 1 illustrates the complementary relationship.



**Figure 1:** Schematic representation of the complementary relationship in regional evapotranspiration (assuming constant energy availability).

### 3. The Regional-Seasonal Advection-Aridity Model

In this section, the components of the complementary relationship— $ET_p$  and  $ET_w$ —are summarized for the Advection-Aridity model. For a more detailed treatment of this model, see *Hobbins et al.* [2001b] and *Brutsaert and Stricker* [1979]. In this model,  $ET_p$  is calculated by combining information from the energy budget and water vapor transfer in the Penman equation, shown below in equation (2), and  $ET_w$  is calculated based on derivations of the concept of equilibrium evapotranspiration under conditions of minimal advection, first proposed by *Priestley and Taylor* [1972], and shown in equation (5).  $ET_a$  is then calculated as a residual of (1).

The familiar expression of the Penman equation for  $ET_p$  is:

$$\lambda ET_p = \frac{\Delta}{\Delta + \gamma} Q_n + \lambda \frac{\gamma}{\Delta + \gamma} E_a \quad (2)$$

where  $\lambda$  represents the latent heat of vaporization,  $\Delta$  is the slope of the saturated vapor pressure curve at air temperature,  $\gamma$  is the psychrometric constant, and  $Q_n$  is the net available energy at the surface. The second term of this combination approach represents the effects of large-scale advection in the mass transfer of water vapor, and takes the form of a scaled factor of an aerodynamic vapor transfer term  $E_a$ .  $E_a$ , also known as the “drying power of the air,” is a product of the vapor pressure deficit and a “wind function” of the speed of the advected air  $f(U_r)$ , of the form (3):

$$E_a = f(U_r) (e_a^* - e_a) \quad (3)$$

In equation (3),  $U_r$  represents the wind speed observed at  $r$  meters above the evaporating surface,  $e_a^*$  and  $e_a$  are the saturation vapor pressure and the actual vapor pressure of the air at  $r$  meters above the surface, respectively.

Previous work by the authors with the Advection-Aridity model [Hobbins *et al.*, 1999, 2001a, 2001b] has shown that model performance is strongly biased with respect to the advective input, and that this bias may be removed by proper selection of the wind function  $f(U_r)$ . In short, the Advection-Aridity model, written and implemented by Brutsaert and Stricker [1979] for use at a temporal scale of three days, and which uses a wind function formulated by Penman [1948], significantly underestimates  $ET_a$  for regions across the conterminous US, generating a mean annual water balance closure error of 7.92% of mean annual precipitation.

The wind function  $f(U_r)$  is either theoretically or empirically derived. Theoretical expressions for the wind function under neutral—i.e., a stable atmospheric boundary layer—conditions are available. While much work has been done in the agricultural arena to calibrate or reformulate wind functions for use in the combination or Penman equation (2) (e.g., Allen [1986], Van Bavel [1966], Wright [1982]), these formulations operate on a limited spatial and temporal scale, do not hypothesize feedbacks of a regional nature, and require local parameterizations of resistance and canopy roughness. Thus, they are not applicable in predicting regional evapotranspiration. Further, in the context of modeling monthly regional  $ET_a$  with the complementary relationship, the effects of atmospheric instability and the onerous data requirements rule out theoretical formulations. In reparameterizing and recalibrating the Advection-Aridity model on a regional, seasonal basis across the conterminous United States, Hobbins *et al.* [2001b] suggest an empirical linear approximation for  $f(U_r)$ , of the form (4):

$$f(U_r) \approx f(U_2)_{i,j} = a_{i,j} + b_{i,j}U_2 \quad (4)$$

where  $U_2$  is the mean monthly wind speed at 2 meters above the ground, and the indices  $i$  and  $j$  represent the month and WRR of application.

The Advection-Aridity model calculates  $ET_w$  using the Priestley and Taylor [1972] partial equilibrium evaporation equation (5):

$$\lambda ET_w = \alpha \frac{\Delta}{\Delta + \gamma} Q_n \quad (5)$$

where the value of the Priestley-Taylor coefficient  $\alpha$  has been the subject of much discussion for various land-cover types and at a range of temporal scales. For example, values of 1.05 [McNaughton and Black, 1973] to 1.32 [Morton, 1983] have been suggested based on short, well-watered Douglas fir and on Morton's [1983] re-examination of the seminal work by Priestley and Taylor, [1972], respectively, while at a range of temporal scales, seasonal values of 1.15 to 1.50 and intra-daily values of between 1.15 and 1.42 have been proposed by DeBruin and Keijman [1979]. Recognizing the variability in estimates for this empirical parameter, its calibration was the focus of much of the work in Hobbins *et al.* [2001b], wherein a value of 1.3177 was shown to remove a consistent bias towards underestimation of  $ET_a$  across a set of 139

basins in the conterminous United States. This value of  $\alpha = 1.3177$  is then used in this study.

Combining the expressions for the wind function  $f(U_r)$  (4), the drying power of the air  $E_a$  (3), potential evapotranspiration  $ET_p$  (2), and wet environment evaporation  $ET_w$  (5) into the expression of the complementary relationship (1), and rearranging yields the following expression for  $ET_a$  (6):

$$\lambda ET_a = (k\alpha - 1) \frac{\Delta}{\Delta + \gamma} Q_n - \lambda \frac{\gamma}{\Delta + \gamma} (a_{i,j} + b_{i,j} U_2) (e_a^* - e_a) \quad (6)$$

where the first term on the right represents the influence of the energy budget, and the second term the local wetness.

In reparameterizing the wind function on a regional and seasonal basis and recalibrating the Priestley-Taylor coefficient  $\alpha$ , *Hobbins et al.* [2001b] reduced the mean annual closure error for large-scale, long-term water budgets to an underestimation of  $ET_a$  to 0.39%. The authors re-examined the applicability of the value of  $k = 2$  for estimating regional  $ET_a$  under non-ideal conditions, and found that in areas of rugged terrain, a value of  $k = 1.946$  may be more applicable. This value is then used across WRRs 14 through 18, i.e., those regions to the west of the Continental Divide.

#### 4. Methodology and Data Sets

This section summarizes the data sets used to generate the  $ET_a$  time series and explains the test used to examine it for temporal trends. To produce the distributed  $ET_a$  time series, the Advection-Aridity model as formulated by the authors requires input data on solar radiation, wind speed, average temperature, humidity, albedo, and elevation. Solar radiation and wind speed data are drawn from “Solar and Meteorological Surface Observation Network (SAMSON)” [NOAA, 1993]. The solar radiation input was the sum of diffuse radiation and topographically corrected direct radiation. Temperature data are drawn from “NCDC Summary of the Day” [EarthInfo, 1998a], and average temperature is estimated as the mean of the average monthly maximum and average monthly minimum temperatures. Humidity data, in the form of dew-point temperatures, are drawn from “NCDC Surface Airways” [EarthInfo, 1998b]. Average monthly albedo surfaces are AVHRR-derived estimates from the *Gutman* [1988] data set. The elevation data were extracted from a 30-arc-second DEM.

In order to generate areal  $ET_a$  coverages, spatial interpolation techniques were applied to the point observations of the input variables. All spatial interpolation and analyses were conducted at a 5-km cell size for the conterminous United States for the period of study (water years 1962-1988). Spatial interpolation procedures are detailed in *Hobbins et al.* [2001a].

At each 5km x 5km cell in the study region, the time series of monthly volumes of  $ET_a$  was examined for trends on an annual basis and on a month-by-month basis. The period under study (water years 1962-1988) lead to one annual series ( $n = 27$ ) and 12 monthly series ( $n = 27$ ).

The presence of trends in the series was tested for using the non-parametric Mann-Kendall test, which is often used to test for trends in

hydrological time series [Lettenmaier *et al.*, 1994]. The Mann-Kendall statistic is given by (7):

$$S = \sum_{k=1}^{n-1} \sum_{j=k+1}^n \text{sgn}(x_j - x_k) \quad (7)$$

where  $n$  is the length of the time series  $x_1 \dots x_n$ , and  $\text{sgn}(\cdot)$  is the sign function. The expected value and variance of  $S$  are given by (8) and (9), respectively:

$$E[S] = 0 \quad (8)$$

$$\text{Var}(S) = [n(n-1)(2n+5) - \sum_t t(t-1)(2t+5)]/18 \quad (9)$$

where the second term represents an adjustment for tied or censored data. The test statistic  $Z$  is then given by (10):

$$Z = \begin{cases} \frac{S-1}{\sqrt{\text{Var}(S)}} & \text{if } S > 0 \\ 0 & \text{if } S = 0 \\ \frac{S+1}{\sqrt{\text{Var}(S)}} & \text{if } S < 0 \end{cases} \quad (10)$$

As a non-parametric test, no assumptions as to the underlying distribution of data are necessary. The test statistic  $Z$  is then used to test the null hypothesis that the data are identically distributed random observations and not time-dependent. The null hypothesis cannot be rejected at the  $\alpha$ -significance level if the following condition (11) applies:

$$|Z| < U_{1-\alpha/2} \quad (11)$$

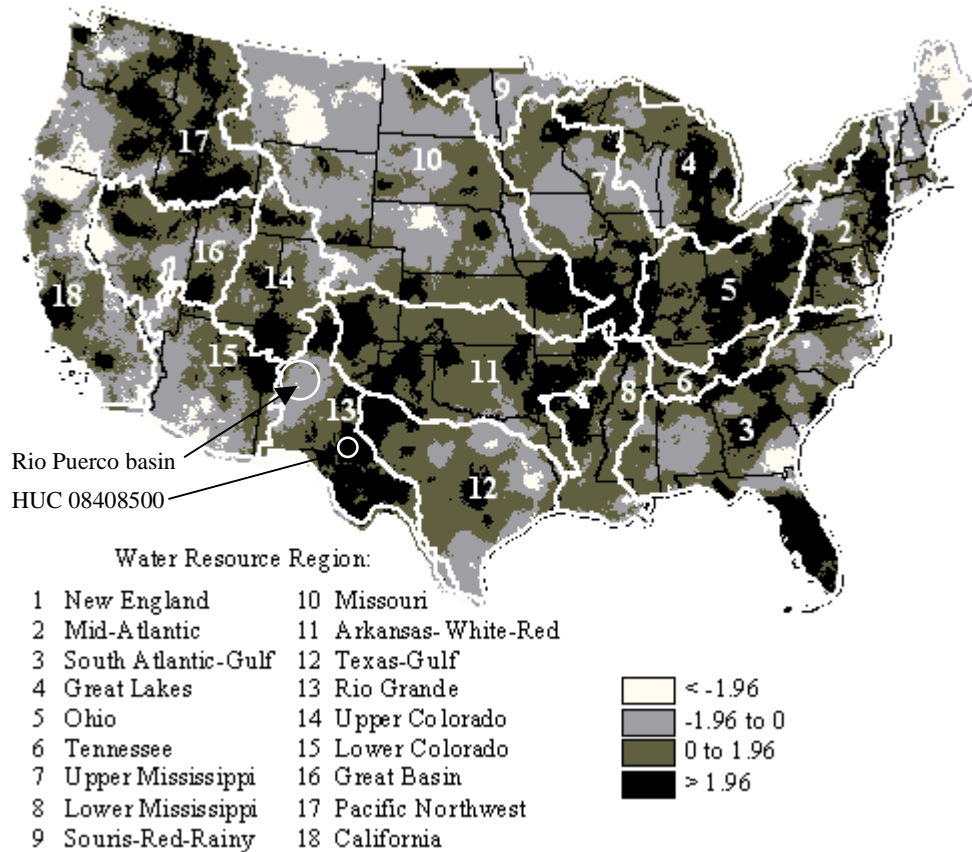
where  $U_{1-\alpha/2}$  is the  $1-\alpha/2$  quantile of the standard normal distribution. Otherwise, the null hypothesis must be rejected at the  $\alpha$ -significance level in favor of the alternative hypothesis of a positive trend if the  $Z$  exceeds the critical value of  $U$  in a positive direction, and vice versa. The confidence level for this study was set at 95%, yielding a critical value of  $|Z|$  of 1.96.

It should be noted that the Mann-Kendall test statistic is non-dimensional, which is to say that it does not offer any quantification of the scale of the trend in the units of the time series under study, but is rather a measure of the correlation of  $x_i$  with time and, as such, simply offers information as to the direction and a measure of the significance of observed trends.

## 5. Results

This paper is not diagnostic, but seeks instead to report the initial findings of an ongoing study. As such, the overall picture of trends in  $ET_a$  across the conterminous United States over the period 1962-1988 are presented; then the trends of the two components of  $ET_a$ —the radiative budget (here represented by  $ET_w$ ) and the mass transfer component (here represented by  $E_a$ )—are examined in the context of the complementary relationship. They are presented spatially across the extent of the study and for a particular sub-region. Finally, a comparison of intra-annual, or seasonal trends with other surface components of the hydrological cycle is made to a basin within that sub-region.

Figure 2 presents the larger picture: that of the distribution of  $ET_a$  trends across the conterminous United States. A preliminary inspection of this figure reveals a complex distribution, with no continental-scale patterns emerging. The exception to this appears to be the Central Lowland Plains and southern

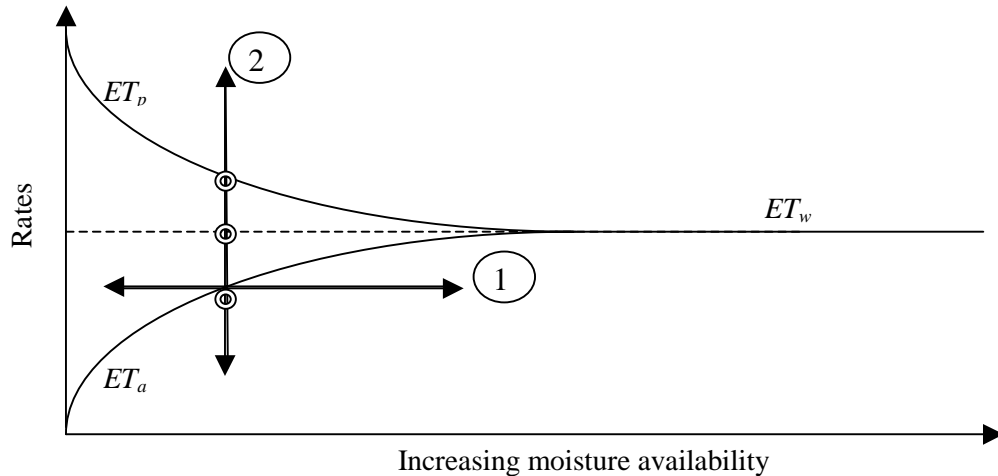


**Figure 2:** Distributed trends in annual  $ET_a$  across the conterminous United States, as measured by the Mann-Kendall test statistic of the time series,  $Z(ET_a)$ . Statistically significant trends at the 95% confidence level have  $|Z(ET_a)| > 1.96$ . Also shown are the boundaries, numbers and names of the WRRs and the approximate locations of the Rio Puerco basin and HUC 08408500.

High Plains where the trend is to increasing  $ET_a$ , and the northern High Plains, where the trend is to decreasing  $ET_a$ . The rest of the study area is a complex, heterogeneous picture from which only a few other generalities can be drawn. WRR 6 (Tennessee), WRR 5 (Ohio) are almost entirely increasing; to lesser degrees, so are WRRs 2 (Mid-Atlantic), 11 (Arkansas-White-Red), 13 (Rio Grande), and 14 (Upper Colorado). WRR 1 (New England) is almost entirely decreasing. The other WRRs are combinations of decreasing and increasing  $ET_a$ , and contain few obvious trends, with the exception of the majority of Florida, which is to increasing  $ET_a$ . An examination of the results at the WRR-scale is somewhat dissatisfying and indicates that this scale may not be optimal for such a discussion.

Having examined the overall picture of trends in  $ET_a$ , attention turns to a first-order determination of what drives these trends and a closer examination of these trends over a particular WRR.

In the context of the complementary relationship, there are two causes of trends in  $ET_a$ : first, by long-term changes in the degree of wetness of the basin, and second by long-term changes in the basinwide radiative budget. For the sake of simplicity of analysis, these two causes may be separated from each other, and their effects are demonstrated in Figure 3.



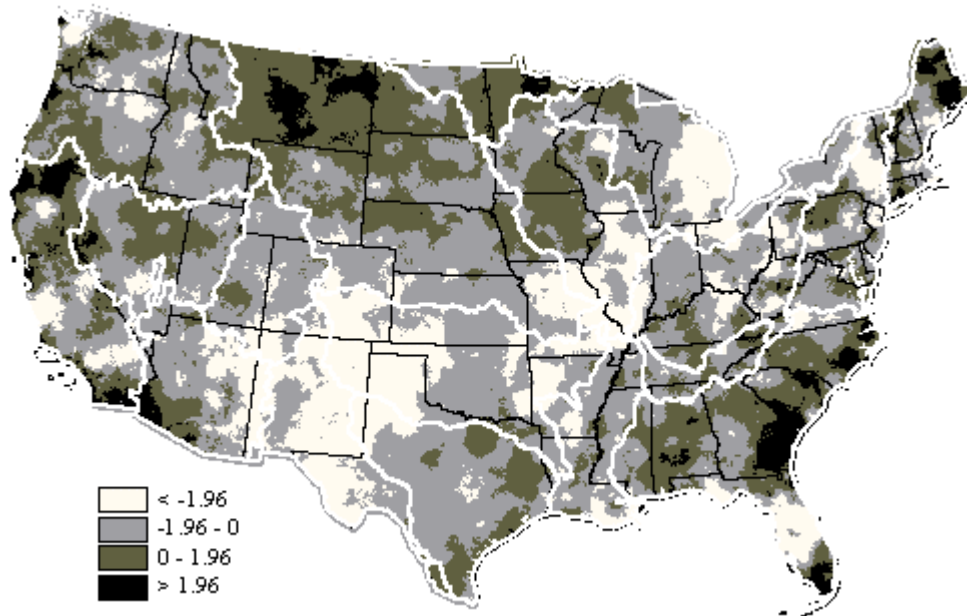
**Figure 3:** The effects of trends in the  $E_a$  and  $ET_w$  components of  $ET_a$  in the context of the complementary relationship. Arrow 1 indicates trends in  $E_a$ , while arrow 2 indicates trends in  $ET_w$ .

The first cause—a long-term change in the wetness of the basin in the absence of a change in the radiative budget—is akin to moving along the paired curves shown in Figure 1. As shown in equation (5),  $ET_w$  is purely a function of the radiative budget  $Q_n$ , and is therefore unaffected by changes in basin wetness alone. According to the complementary relationship and as shown in Figure 3, increasing wetness moves the  $ET_a$ - $ET_p$  pair to the right, towards convergence of their respective curves, while decreasing basin wetness moves the  $ET_a$ - $ET_p$  pair divergently to the left. This is shown by the arrow marked ‘1’ in Figure 3. This dynamic may be indicated by the Mann-Kendall test statistic for the  $E_a$  time series,  $Z(E_a)$ . If, for a particular basin,  $Z(E_a)$  is negative, then the basin is becoming wetter from the first effect; drier for a positive  $Z(E_a)$ . The spatial distribution of  $Z(E_a)$  is shown in Figure 4.

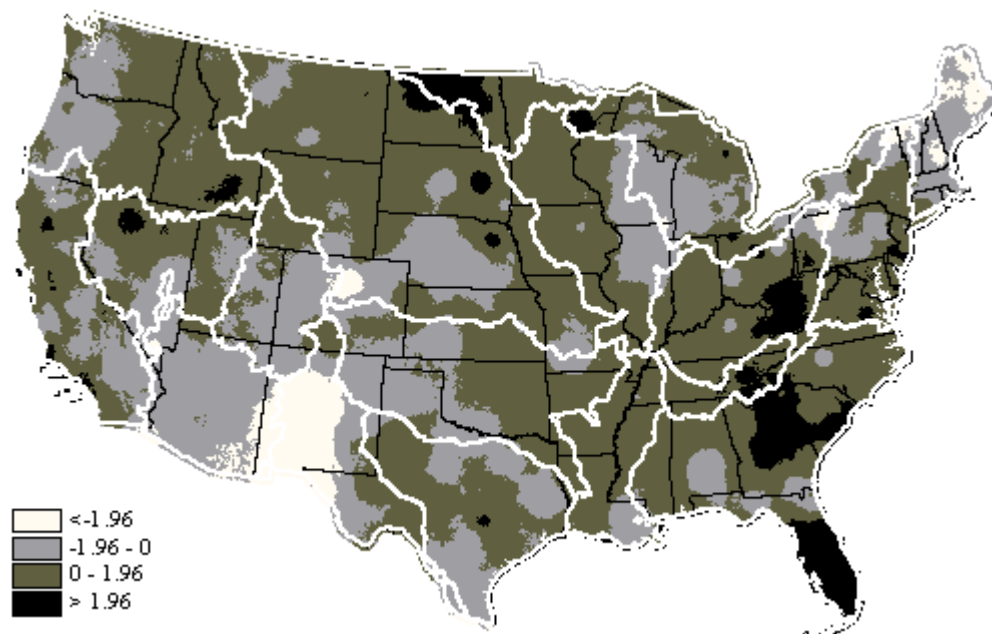
The second cause—a change in the long-term radiative budget in the absence of a change in wetness—shifts both the straight line representing  $ET_w$  and the curve representing  $ET_p$  upwards, in the case of increasing  $Q_n$ , or downwards in the case of decreasing  $Q_n$ . Correspondingly, and in accordance with the complementary relationship, the curve representing  $ET_a$  will also shift in a similar direction, as shown by the arrow marked ‘2’ in Figure 3. As  $ET_w$  is a function of  $Q_n$  (see equation (5)) trends in the radiative budget may be tracked with the value of the Mann-Kendall test statistic for  $ET_w$ ,  $Z(ET_w)$ . The spatial distribution of  $Z(ET_w)$  is shown in Figure 5.

It is important to recognize that, physically, these dynamics cannot be taken in isolation, but instead occur within the framework of complex

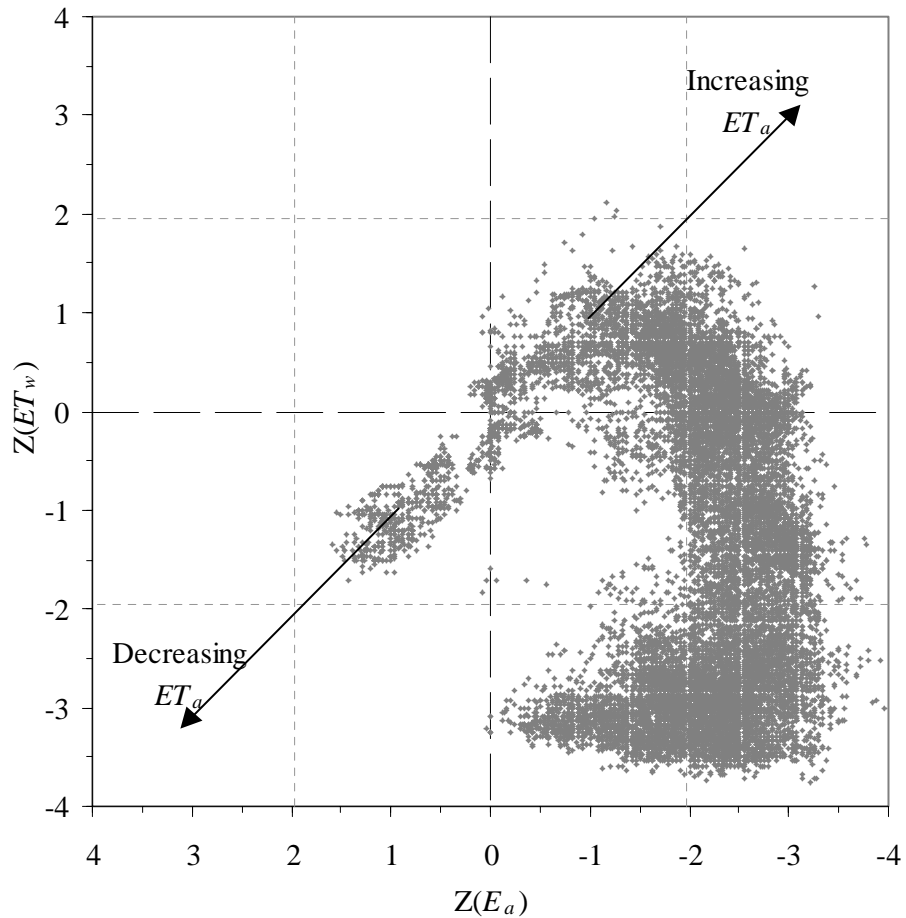
feedback mechanisms beyond those that underpin the complementary relationship as it is currently formulated. For instance, the effect of a long-term decrease in the radiative budget, indicated here by a negative  $Z(ET_w)$  value, may be to increase soil moisture and thereby to decrease the vapor pressure deficit of the over-passing air, which leads, in turn, to a lowering of the drying power of the air, and hence a negative  $Z(E_a)$ .



**Figure 4:** Distributed trends in annual  $E_a$  across the conterminous United States, as measured by the Mann-Kendall test statistic of the time series,  $Z(E_a)$ . Statistically significant trends at the 95% confidence level have  $|Z(E_a)| > 1.96$ . Also shown are the boundaries of the WRRs.



**Figure 5:** Distributed trends in annual  $ET_w$  across the conterminous United States, as measured by the Mann-Kendall test statistic of the time series,  $Z(ET_w)$ . Statistically



significant trends at the 95% confidence level have  $|Z(ET_w)| > 1.96$ . Also shown are the boundaries of the WRRs.

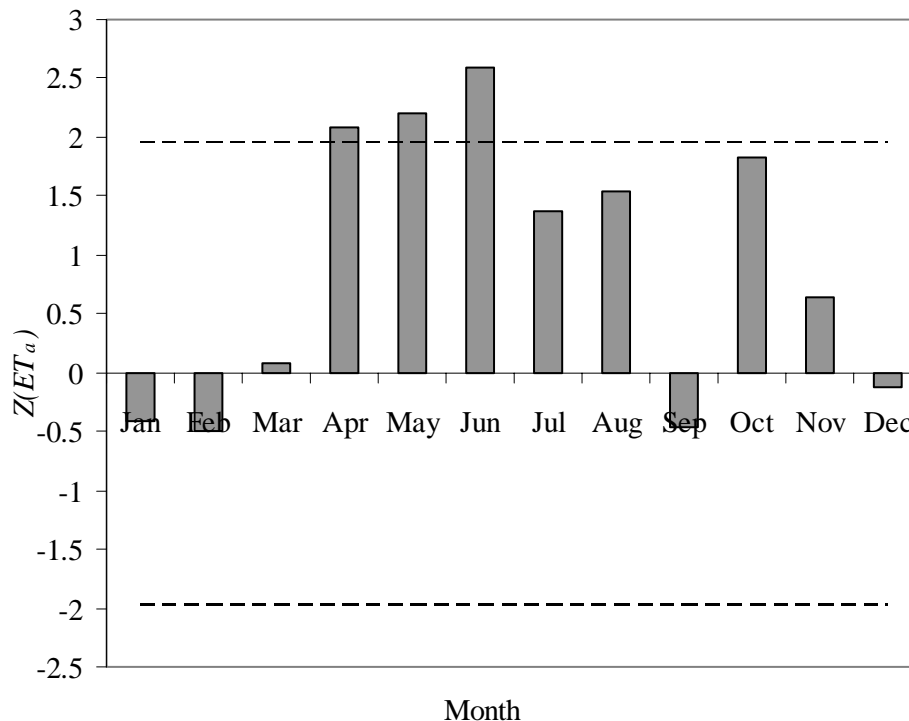
**Figure 6:** Mann-Kendall test statistics for  $ET_w$  and  $E_a$  components of  $ET_a$  for all cells in WRR 13 (Rio Grande). Statistics are plotted orthogonally, with  $Z(E_a)$  plotted in reverse (positive left, negative right). Statistically significant trends at the 95% confidence level have  $|Z(ET_w)|$  or  $|Z(E_a)| > 1.96$ .

Figure 6 indicates, for each 5km x 5km cell in WRR 13 (the Rio Grande basin), the relative trends in the driving factors behind trends in  $ET_a$ . In this figure, the y-axis presents  $Z(ET_w)$ , which here represents any trend in the radiative budget; increasing above the x-axis and decreasing below. The x-axis presents  $Z(E_a)$ , which here represents trends in basin wetness; increasing  $E_a$  and hence decreasing  $ET_a$  to the right of the y-axis, and decreasing  $E_a$  and hence increasing  $ET_a$  to the left. The axes are shown in this manner in order to correspond to the trends indicated in Figure 3, wherein the first cause for a trend in  $ET_a$ —a trend in the radiative budget—moves a cell horizontally, and the second cause—a trend in wetness moves a cell in the vertical direction.

Thus, each quadrant of Figure 6 represents a different combination of influences driving the  $ET_a$  trends. Cells in the upper-right quadrant are experiencing trends in  $ET_w$  and  $E_a$ —increasing in the former and decreasing in the latter—that both tend to increase  $ET_a$ . These cells are indicated as positive cells in Figure 5, negative cells in Figure 4, and positive cells in Figure 2. Cells in the lower-left quadrant correspond to the opposite effects in both  $ET_w$  and  $E_a$  and thus indicate a decreasing trend in  $ET_a$  and are shown as negative cells in Figure 2. The other two quadrants indicate mixed effects on  $ET_a$ , when  $ET_w$  and  $E_a$  are both increasing or are both decreasing.

The spatial distribution of these orthogonal contributing trends can be derived from Figures 4 and 5 and their effects in Figure 2. Those cells that, from Figure 6, we can predict are decreasing in  $ET_a$  are the areas along the lower reaches of the basin, in southern Texas. The cells in the top-right quadrant lie in the far northern reaches of the WRR – the San Luis Valley, eastern San Juan and western Sangre de Cristo mountain ranges in southern Colorado, and in western Texas and far southeastern New Mexico. The rest of the WRR is subject to conflicting trends in these two factors: either drying but receiving more solar radiation, or wetting but receiving less solar radiation.

Work by *Molnár and Ramírez* [2001] on recent trends in precipitation and streamflow in the Rio Puerco basin, which is located in west-central New Mexico in the western reaches of WRR 13 (Rio Grande), allows us a coarse comparison with the other components of the surface water budget. The monthly trends in  $ET_a$  over the 27-year period of this study for an 8-digit HUC (08408500) in this WRR are shown in Figure 7.



**Figure 7:** Mann-Kendall test statistics for monthly  $ET_a$  time series for HUC 08408500 in WRR 13 (Rio Grande). Statistically significant trends at the 95% confidence level have  $|Z(ET_a)| > 1.96$ .

On a seasonal basis at HUC 08408500, the increasing  $ET_a$  trends are distributed throughout the spring and early summer months (April through August) and in early fall (October and November).  $ET_a$  appears to be decreasing, though not significantly, through the winter months (December through February). The only trends that are significant at the 95% confidence level are those for April through June. This seasonal pattern does not correspond to the seasonal patterns reported for either precipitation or streamflow in the Rio Puerco basin, where both variables increase from September to May and decrease in the summer months, with the only significant trends reported being those for streamflow from November through May.

*Molnár and Ramírez* [2001] report an increase in annual precipitation at the Rio Puerco basin from ~210mm in the 1950's to ~340mm in the 1990's, with an increase in wetness in the non-summer months and no corresponding significant trends in annual streamflow. This implies either an increase in basin storage or a significant increase in  $ET_a$ . While *Molnár and Ramírez* [2001] had no climatologically derived or directly observed  $ET_a$  data to test this implication, examination of the area of their study in Figure 2 indicates that, as estimated by the Advection-Aridity model, annual  $ET_a$  at the Rio Puerco basin has decreased, though not significantly. At HUC 08408500, however, annual  $ET_a$  shows a significant increase.

Potential reasons for the discrepancy between the two results for the Rio Puerco basin are many-fold. Firstly, it must be considered that the temporal extents of the two studies are not identical: 35 years—from 1952 to 1986—for *Molnár and Ramírez* [2001] and 27 years—from 1962 to 1988—for this study. Also, the basins may simply be too far removed from each other for adequate comparison. Indeed, they lie approximately 150 miles across mountain ranges from each other. That there seems no correlation between the two basins raises again the possibility that examining trends on this scale is meaningless, especially in such heterogeneous terrain.

Further, comparing trends in precipitation and streamflow to trends in  $ET_a$  is difficult at best, especially in light of the fact that the  $Z$ -statistics are not quantifications of trends, but simply measures of trends' direction and significance. Instead, it may be of more use to generate a lumped, or basinwide, time series of annual  $ET_a$  estimates for the Rio Puerco basin, based on the difference between precipitation and streamflow, and then to compare this quantity to the annual modeled  $ET_a$  time series at the 8-digit HUC. However, this would not allow for seasonal trend data to be derived.

## 6. Conclusions

This paper reports on the initial findings of a study wherein a spatially distributed 27-year monthly time series of  $ET_a$  for the conterminous United States was produced, and outlined a procedure whereby temporal trends in this data set could be examined at different spatial and temporal scales.

Distributed surfaces of these trends were produced and examined, firstly over the spatial extent of the data set and then over a particular sub-region within it. Further, for a watershed within this sub-region, intra-annual trends were derived and these results were compared to work on precipitation and streamflow in a nearby basin in the same region.

While the trends were derived and categorized as to their origins within the context of the driving forces of  $ET_a$ , no attempt was made to explain these trends on a climatological basis. Nor were other models—such as ecological, land-use, or general circulation models—examined for similarity of results. This type of diagnostic analysis is the subject of ongoing study.

Work is underway to examine the  $ET_a$  trends with respect to the input data—specifically temperature—and to other components of the surface hydrological cycle—namely precipitation and streamflow. The comparison of trends in modeled  $ET_a$  at one basin to previously reported trends in precipitation and streamflow at another was disappointing for various reasons involving non-contemporaneity of temporal extent, spatial dislocation of the basins, and the difficulties inherent in comparing trends in different hydrological variables. Resolution of these difficulties will allow comparison of the  $ET_a$  trends within the context of trends that are readily observable and that have been established within the existing science of hydrology. Further, comparisons of the  $ET_a$  trend data should be directly comparable to data from other branches of the science—models of land-use, ecological dynamics, or global circulation, for instance.

Other work on trends in  $ET_a$  will also be examined, as and when it becomes available.

With regard to the specific methodologies outlined in this paper, it may be concluded that dividing  $ET_a$  trends up as to their origins may be a useful tool to determine effects of climatological trends on  $ET_a$ . Further, dividing spatially according to WRR may be too coarse a taxonomy. For instance, WRR 11 (Missouri) includes the Rocky Mountains at the Continental Divide on the Idaho-Montana border, vast tracts of the High Plains, and extends as far south-east as the Mississippi River on the Missouri-Illinois border. Examining results from such a large and diverse single region may not be a useful tool and may instead have the effect of rendering useful data noise. This conclusion is supported by Lins [1997] who, in a study of regional streamflow patterns and hydroclimatology, concluded that while the division of the conterminous United States into the 18 WRRs as promulgated in 1970 by the Water Resources Council is based purely on surface watersheds, they do not correspond to observed hydroclimatology, as many of the hydrological driving forces are not constrained by the drainage divides that define the WRRs. Another, improved taxonomy may be derived by sub-dividing the WRRs into their component 4-digit HUCs, or on the 344 state climatic divisions.

A potential drawback to the methodologies described in this paper—specifically the use of the regional, seasonal Advection-Aridity model—should here be addressed. The reparameterization of the Advection-Aridity model—i.e., recalibrating the values for  $\alpha$ ,  $k$ , and the parameters in the regional, seasonal  $f(U_2)$ - $U_2$  relationships—was conducted under the

assumption that, in the long-term and on a basin-scale, the sum of  $ET_a$  and streamflow equals precipitation, and therefore that net change in basin storage over the time-extent of the study is zero. In the context of climate change studies, however, this assumption may no longer be tenable. Trends in precipitation, streamflow, and even  $ET_a$  may, in fact, lead to an increase or decrease in basin storage. The Advection-Aridity model used to generate the  $ET_a$  data for this study has been reparameterized such that, should any such trends underlie the hydrological components, they are eliminated on a mean basis and over the length of the record, that is to say that the mean (or rather, the volume-weighted mean) value of the trend is subtracted. To use a model that has been calibrated under these circumstances to examine trends in the data used to calibrate the model will not introduce spurious trends, but will instead have the effect of mediating existing trends. Trends that may be significant in the input variables to the model (in temperature, or  $Q_n$ , for example) may be rendered insignificant by the filtering effect of the calibrated parameters when the inputs are applied to generate an output ( $ET_a$ ) from the model. It should be noted, however, that the validity of trends examined in this paper is not called into question by this procedure, rather the magnitude of any existing trends may be underestimated.

Using the methodologies summarized in this paper, it is possible to examine trends in  $ET_a$  and its components in any and all basins across the conterminous United States, as the input data are distributed and reflect surface conditions regardless of the origins of the water and degree of anthropogenic disturbance. Thus, after completion of the further work described herein, important information about a heretofore missing link in surface hydrology can be made available to water managers, climatologists, and ecologists, or to any student of the surface hydrological cycle.

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