Modeled Areal Evaporation Trends over the Conterminous United States

By Jozsef Szilagyi[1]

Abstract: Long-term (1961–1990) areal evapotranspiration (*AE*) has been modeled with the help of 210 stations of the Solar and Meteorological Surface Observation Network within the conterminous United States. Modeled *AE*, averaged over all stations, has shown an overall increase of about 2–3% in the period 1961–1990, both on an annual basis and over the warm season (May–September). The rate of increase has differed among three geographic regions: the eastern, central, and western United States, with the largest modeled increase found in the east, followed by the central part of the United States. In the western part of the continent, modeled *AE* has, in fact, stayed constant. Of these trends, only the ones over the eastern part of the conterminous United States are statistically significant.

INTRODUCTION

It is widely accepted today that any change in the climate system will have important consequences for water resources management and conservation since more and more people depend on a resource already scarce in many parts of the world (e.g., the semidesert and desert parts of the United States). As humanity alters the Earth's environment on a global scale, it also interacts with the global hydrologic cycle (Vorosmarty et al. 2000); however, to what extent is still largely unknown. While much effort is directed toward predicting the future climate of the Earth, there are still gaps in our understanding of today's status of the hydrologic cycle. The influence of such hydrologic processes as evapotranspiration on the Earth's climate is increasingly seen as particularly significant (Committee on Opportunities in the Hydrologic Sciences 1991). To better prepare ourselves for future changes in water resources management and conservation, we must learn about how evapotranspiration changes on a long-term basis. In long-term water resources planning and management, trends in evapotranspiration must be quantified, since it is a consumptive water use that cannot be recovered (Committee on Opportunities in the Hydrologic Sciences 1991). This present study investigates whether any such trend in areal evapotranspiration (*AE*) is detectable over the conterminous United States.

Intuitively, an increase in long-term mean annual temperatures over the contiguous United States (Karl et al. 1996), in conjunction with an increase in long-term precipitation (Lettenmaier et al. 1994; Karl et al. 1996), would be expected to translate into enhanced runoff and *AE* values. While runoff has indeed been documented as increasing (Lettenmaier et al. 1994; Lins and Slack 1999) over the conterminous United States, the same confirmation has yet been missing for *AE*, since this latter value cannot be simply measured.

The most straightforward estimation of *AE* ($M\ L^{-2}\ t^{-1}$) on an annual basis comes from the water-balance equation applied over a watershed

$$AE = P - RO \tag{1}$$

where P = precipitation ($M\ L^{-2}\ t^{-1}$); and RO = runoff ($M\ L^{-2}\ t^{-1}$). In (1) it is assumed that (1) no significant changes occur in water storage between hydrologic years (i.e., October–September); (2) there is no other source of ground water in the catchment than recharge from precipitation; and (3) the only ground-water sink is via base flow to the stream where runoff is measured. The obvious difficulty with this technique is that the assumptions are almost impossible to verify. The accuracy of the *AE* estimates is generally unknown, not only due to the above factors but also due to possibly significant measurement uncertainties in both the precipitation and discharge measurements.

Recently, there has been a renewed interest (Katul and Parlange 1992; Parlange and Katul 1992; Kim and Entekhabi 1997; Brutsaert and Parlange 1998) in the application of a technique, called Bouchet's complementary hypothesis (Bouchet 1963), to estimate *AE*. In the present study this approach, as further developed by Morton (1983) and Morton et al. (1985), is employed to investigate trends in *AE* over the conterminous United States. Hobbins et al. (1999) pointed out recently that Morton's WREVAP model (Morton 1985), an upgraded version of his own complementary relationship areal evapotranspiration (CRAE) model (Morton 1983), performs better in estimating annual watershed *AE* than the often used advection aridity (Brutsaert and Stricker 1979) approach.

Below it will first be demonstrated that the complementary hypothesis is valid with the help of satellite-derived energy-balance data. Then it will be shown how Morton's WREVAP model outputs compare with the satellite-derived energy-balance data, as well as with pan-evaporation measurements. Finally trends in annual and warm-season (May–September) *AE* in three regions (east, central, and west) within the United States and over the entire conterminous United States will be estimated.

THEORY

Starting with Bouchet's complementary hypothesis (Bouchet 1963), *AE* is related to potential [E_p ($M\ L^{-2}\ t^{-1}$)] and wet surface [E_w ($M\ L^{-2}\ t^{-1}$)] evaporation via

$$AE = 2E_w - E_p \tag{2}$$

where E_p, for example, can be estimated using pan-evaporation data (Brutsaert and Parlange 1998) or a Penman-like (Penman 1948) combination equation (Katul and Parlange 1992; Parlange and Katul 1992). In (2) E_w is the evaporation "that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water" (Morton 1983). Thus E_w is limited mainly by the available energy (Bouchet 1963) at a given location. For example, in the approach of Priestly and Taylor (1972)

$$L \cdot E_w = \alpha \frac{\delta}{\delta + \gamma} R_n \tag{3}$$

where α = constant; δ = slope of the saturation vapor pressure-temperature curve ($M\ t^{-2}\ L^{-1}\ T^{-1}$); γ = psychrometric constant

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(M t^{-2} L^{-1} T^{-1}); and L = latent heat of vaporization (E M^{-1}). The $\delta/(\delta + \gamma)$ term changes only slightly with temperature (Brutsaert 1982), which is a function of R_n.

According to (2), AE changes with respect to changes in E_p and/or E_w. The idea behind (2) can be demonstrated by comparing a desert environment, where E_p is extremely high and where AE is extremely low, with a more humid environment, where AE is almost as high as E_p, which, in turn, is much smaller than in the desert case.

There are several assumptions (Bouchet 1963; Morton 1983) to be made when deriving (2). While in practice, when applying (2), these assumptions may not be strictly upheld; the hypothesis itself can be validated not only on an annual basis, including water balances (Morton 1983; Hobbins et al. 1999), but on a monthly basis as well, using the energy balance.

Validation of Bouchet's Hypothesis

The net energy [R_n(E L^{-2} t^{-1})] that the vegetated surface receives is used by the sensible-heat [H(E L^{-2} t^{-1})] and latent-heat ($L \cdot AE$) turbulent exchanges between the surface and the air and by heat conduction [G(E L^{-2} t^{-1})] within the surface (both vegetation and soil)

$$R_n = H + L \cdot AE + G \tag{4}$$

Choosing a monthly time-increment and considering adjacent months when R_n stays constant, (4) transforms into

$$L \cdot \Delta AE = -\Delta H \tag{5}$$

where Δ designates the change in the monthly average value of a variable between months, and where it was assumed that the change in surface heat conduction between months with no change in R_n, is negligible. Rewriting (2) for such months one obtains

$$\Delta AE = -\Delta E_p \tag{6}$$

since the E_w terms is primarily a function of R_n only. Combining (5) and (6) one obtains

$$L \cdot \Delta E_p = \Delta H \tag{7}$$

The change in the sensible-heat term can be expressed (Dingman 1994) as

$$\Delta H = K_H \cdot \Delta(u \cdot TD) + (u \cdot TD) \cdot \Delta K_H \tag{8}$$

where K_H = turbulent transfer coefficient for sensible heat (E T^{-1} L^{-3}); u = wind velocity (L t^{-1}); and TD = temperature difference between the surface and the air (T). The left-hand side of (7) can be written as

$$L \cdot \Delta E_P = L \left\{ \frac{\gamma}{\delta + \gamma} \Delta[K_E \cdot u \cdot VPD] \right\} \tag{9}$$

where the terms within the outermost braces come from the Penman equation (Penman 1948)

$$L \cdot E_P = \frac{\delta}{\delta + \gamma} R_n + \frac{\gamma}{\delta + \gamma} L \cdot E_A \tag{10}$$

when R_n is constant between adjacent months. Here again the assumption was used that the term involving the psychrometric constant would change only negligibly between such months. In (10) K_E is the turbulent transfer coefficient for latent heat (t^{-2} L^{-2}); E_A (=$K_E \cdot u \cdot VPD$) is called the drying power of the air (Katul and Parlange 1992); and VPD is the vapor pressure deficit (M t^{-2} L^{-1}) (i.e., the difference between saturation and actual vapor pressure). Assuming near-neutral atmospheric stability conditions (justified by monthly averaging and the close proximity of standard meteorological measurements to the surface), and also assuming negligible change in the transfer coefficients between months with no change in R_n, one can write

$$1 = \frac{\Delta H}{L \cdot \Delta E_P} \approx \frac{K_H \cdot \Delta(\langle u \rangle \cdot \langle TD \rangle)}{L \dfrac{\gamma}{\delta + \gamma} K_E \cdot \Delta(\langle u \rangle \cdot \langle VPD \rangle)} = (\delta + \gamma) \frac{\Delta(\langle u \rangle \cdot \langle TD \rangle)}{\Delta(\langle u \rangle \cdot \langle VPD \rangle)} \tag{11}$$

where the following identity (Dingman 1994) was employed

$$\frac{K_H}{L \cdot K_E} = \frac{c_a}{L \left(\dfrac{0.622}{P_a} \right)} = \frac{c_a P_a}{0.622 L} = \gamma \tag{12}$$

where c_a = heat capacity of air (E M^{-1} T^{-1}); and P_a = its pressure (M t^{-2} L^{-1}). The parentheses represent monthly average values. For the surface temperature only monthly mean values were available; consequently the average of the daily horizontal wind velocities had to be taken first, followed by the multiplication. To be consistent, the same was done for the term involving E_p. In general, however, $\langle x \cdot y \rangle = \langle x \rangle \cdot \langle y \rangle$ only if x and y are uncorrelated. Note that due to the above monthly averaging the H term may significantly be underestimated (Bodyko 1974, p. 91; Brutsaert 1982, p. 208; Mintz and Walker 1993) even if only the change in the H (and E_p) variable between adjacent months is needed. This must be kept in mind when Bouchet's hypothesis is tested in practice below.

If the quantity on the right-hand side of (11) can be shown to equal unity when R_n is constant between adjacent months, then Bouchet's hypothesis has been validated. Note that this does not mean that Bouchet's hypothesis is valid only when R_n is constant. A constant R_n term is not a requirement for Bouchet's hypothesis. Since the hypothesis has been shown to be valid either on an annual basis using water balances (Morton 1983; Hobbins et al. 1999) or locally, over short time periods (i.e., hours) using point measurements (Parlange and Katul 1992; Kim and Entekhabi 1997), it was thought worthwhile to show that it also works over intermediate time (i.e., months) and regional spatial scales (i.e., ~300 km, the scale of satellite-derived radiation and surface temperature measurements) required by the WREVAP model.

To test this, the 8-year surface radiation budget (SRB) monthly dataset of the Langley Distributed Active Archive Center has been used in conjunction with daily meteorological measurements [averaged over each month between 1984 and 1991, i.e., the temporal coverage of the SRB data] of 132 National Weather Service Stations (Fig. 1) and distributed by the National Climatic Data Center (NCDC). The SRB dataset provided the monthly mean R_n and surface temperature values

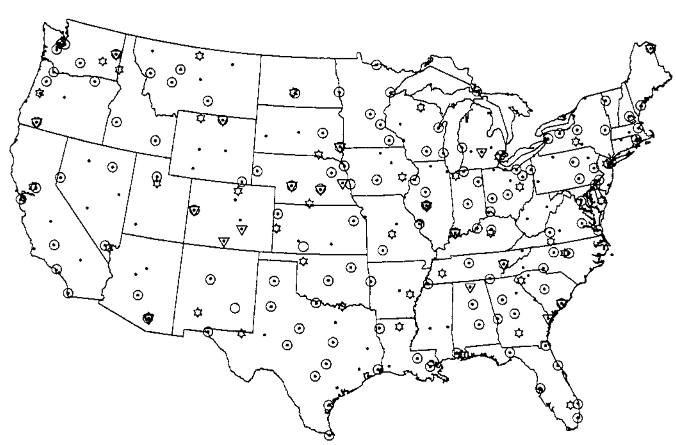

FIG. 1. Distribution of: a) SAMSON Stations (Dots); (b) NCDC Stations (Circles); (c) SAMSON Stations with Long-Term Pan-Evaporation Data (Triangles); (d) NCDC Stations with Long-Term Pan-Evaporation Data (Stars)

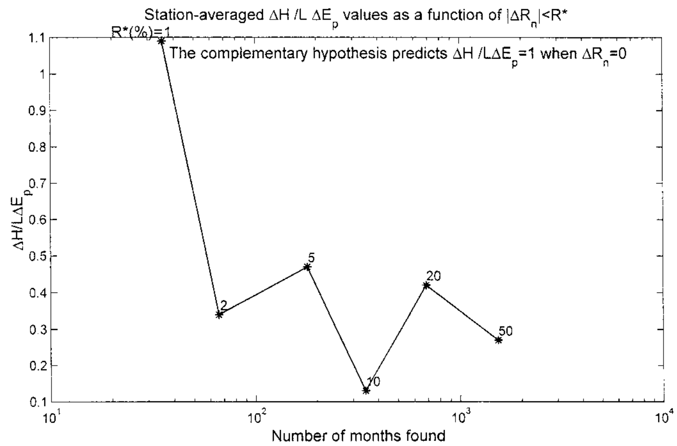

FIG. 2. Validation of Complementary Hypothesis Using NCDC and SRB Data

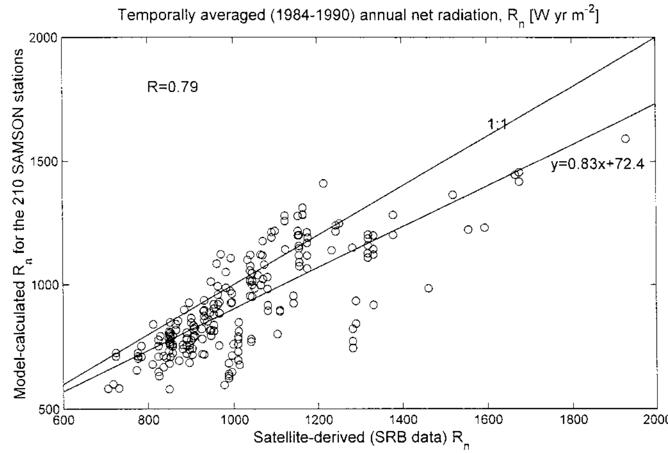

FIG. 3. Satellite-Derived and WREVAP-Estimated R_n Values

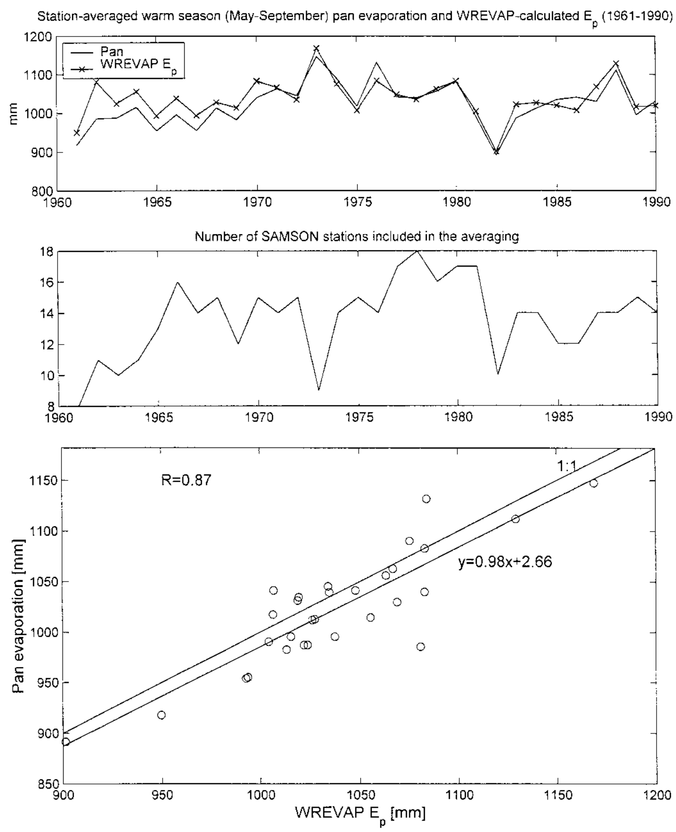

FIG. 4. Warm-Season Pan-Evaporation Sums and WREVAP E_p Estimates

(for the calculation of ΔH), representative over grid cells with dimensions of about 280 by 280 km. For each NCDC station the monthly mean ΔH and δE_p values were calculated whenever ΔR_n was smaller than a predefined value (R^*) in adjacent months over the grid cell in which the actual station fell. Fig. 2 displays the station-averaged $\delta H/L\Delta E_p$ values as a function of R^*. The abscissa indicates the number of occasions when $|\Delta R_n| < R^*$ was observed. As ΔR_n approaches zero, the ratio significantly shot up, in accordance with the complementary hypothesis. The exact values of the ratios, however, should be treated with extreme caution due to the above discussed reasons.

APPLICATION OF WREVAP MODEL

Morton's WREVAP model (Morton et al. 1985) calculates *AE* using Bouchet's complementary hypothesis (Bouchet 1963). It has been tested extensively by Morton over 143 river basins in four continents and gave very close estimates of annual *AE* (Morton 1983) in comparison with (1) applied over the watersheds. Recently, Hobbins et al. (1999) corroborated the reliability of the WREVAP model-calculated annual *AE* estimates over 362 catchments that are only minimally affected by interbasin and intrabasin water diversions and/or groundwater pumping.

For the present purpose, the model was run on a monthly basis with inputs such as mean monthly temperature and mean monthly dew-point temperature, as well as monthly sums of incident global radiation [R_s (E L^{-2} t^{-1})]; all obtained from the Solar and Meteorological Surface Observation Network (SAMSON) dataset that has 210 stations (Fig. 1) within the conterminous United States for the period 1961–1990. Since the calculation of *AE* includes the estimation of R_n, it was checked how the WREVAP-estimated R_n values compare with the satellite-derived (SRB) values. Fig. 3 shows that the WREVAP-estimated multiyear mean annual net radiation values for the 210 SAMSON stations are comparable with the SRB pixel values (the correlation coefficient is 0.79); however, WREVAP tends to slightly undershoot R_n.

The next check compared the WREVAP-calculated E_p estimates with pan evaporation measurements. There are 19 SAMSON stations (Fig. 1) with long-term pan-evaporation data. Fig. 4 shows how close the WREVAP estimates are to the measured values, with a correlation coefficient of 0.87 for the station-averaged warm-season accumulates of the monthly sums.

Based on these tests, plus the annual water-balance comparisons published by Morton (1983) and Hobbins et al. (1999), it was concluded that the WREVAP model may be expected to give reliable estimates of long-term *AE* trends.

RESULTS AND DISCUSSION

The WREVAP model was run on a monthly basis over the period 1961–1990, the temporal coverage of the SAMSON data. The model-estimated time series of the E_p, E_w, and *AE* values are displayed in Figs. 5–8 for the entire conterminous United States and for three subregions (eastern, central, and western United States, respectively). The results are further divided into annually and warm-season-representative values. A linear trend-function was fit to each time series, followed by a linear trend-analysis (using t-tests) in each case (Table 1). A common feature of all time series in Figs. 5–8 is that the trend functions always display a nonnegative slope. However, based on the trend-analysis here (Table 1), only the wet-surface (E_w) and areal evapotranspiration (*AE*) trends show a statistically significant increase. The former is over the entire conterminous United States, as well as in two subregions (the eastern and western United States), while the latter one is only in the eastern part of the conterminous United States. An increase in the E_w estimates is probably a result of increasing temperatures (Karl et al. 1996) and a possible increase in the net energy (R_n) term. While this latter trend is not statistically significant, the estimated long-term changes are almost always positive (Table 1). In Table 1, in addition to the t-test, a second, rather arbitrary, test was applied as well. The slope in the linear trend-function was deemed to represent a trend (and marked it by ''Y'' in Table 1) if the relative change [d_r(%)] was larger than 2.5% [i.e., the difference in the function values evaluated in the last (1990) and first year (1961) of the period and divided by the mean value of the function over the period].

An increase in modeled *AE* is a straight consequence of an increase in E_w and near-constant E_p values (Table 1), according to the complementary hypothesis [(2)]. While long-term pan-evaporation values have been reported to decline (Peterson et al. 1995) in the United States over the past 50-some years, the long-term pan-evaporation data available for 40 NCDC sites (Fig. 1) express a practically constant long-term mean (Fig. 9), in accordance with modeled near-constant long-term E_p.

In summary, it can be stated that the WREVAP model estimates of monthly areal evapotranspiration values for the period 1961–1990 express a statistically significant 4% relative increase over the eastern part of the conterminous United States. A 4.5% relative increase in warm-season (May–September) areal evapotranspiration estimates for the same region has also been reported here. The same values for the conterminous United States are 2.5 and 3%, respectively; however, neither of them is statistically significant. Increased levels of

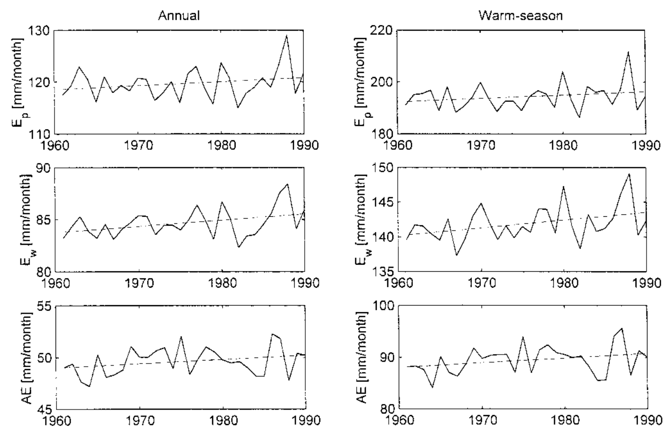

FIG. 5. Station-Averaged (210 SAMSON Stations) Annual and Warm-Season E_p, E_w, and *EA* Estimates with Their Linear Trend Functions, Conterminous United States

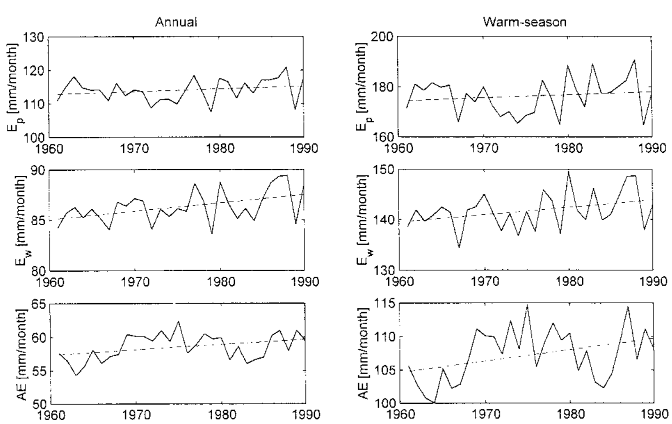

FIG. 6. Station-Averaged (77 SAMSON Stations) Annual and Warm-Season E_p, E_w, and *EA* Estimates with Their Linear Trend Functions, Eastern Conterminous United States

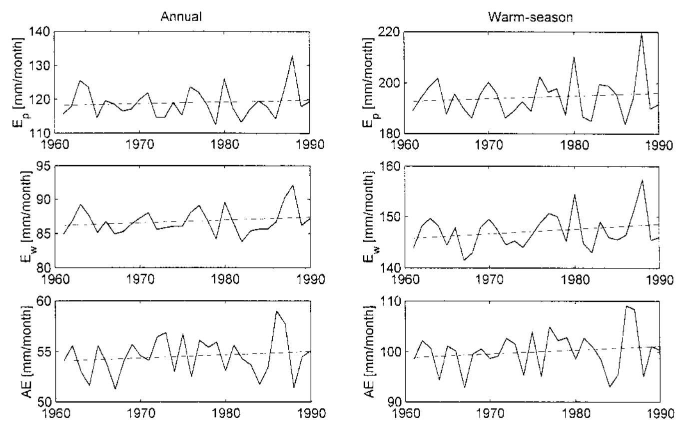

FIG. 7. Station-Averaged (79 SAMSON Stations) Annual and Warm-Season E_p, E_w, and *EA* Estimates with Their Linear Trend Functions, Central Conterminous United States

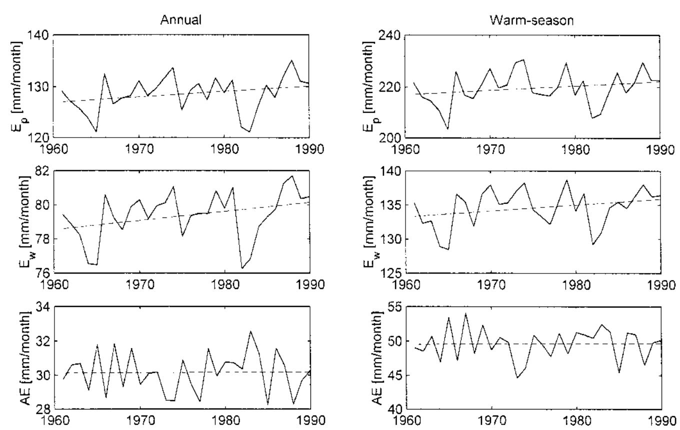

FIG. 8. Station-Averaged (55 SAMSON Stations) Annual and Warm-Season E_p, E_w, and *EA* Estimates with Their Linear Trend Functions, Western Conterminous United States

TABLE 1. Results of t-Tests (d^*) for Null-Hypothesis (H_0): Time Series Expresses Trend

Parameter		U.S. annual	U.S. warm-season	Eastern-U.S. annual	Eastern-U.S. warm-season	Central-U.S. annual	Central-U.S. warm-season	Western-U.S. annual	Western-U.S. warm-season
R_s	d^*	No	No	No	No	No	No	No	No
	d_r	0.00	0.41	0.25	0.92	−0.26	0.18	−0.14	0.00
R_n	d^*	No	No	No	No	No	No	No	No
	d_r	0.88	1.19	1.70	2.17	0.78	0.99	−0.29	0.00
E_w	d^*	Y_{95}	Y_{95}	Y_{95}	Y_{90}	No	No	Y_{90}	No
	d_r	2.04	2.28	2.76 y	2.97 y	1.38	1.86	1.95	1.90
E_p	d^*	No	No	No	No	No	No	No	No
	d_r	1.88	1.96	2.18	1.99	1.22	1.69	2.38	2.27
AE	d^*	No	No	Y_{90}	Y_{95}	No	No	No	No
	d_r	2.41	2.97 y	3.87 y	4.59 y	1.73	2.20	0.16	0.24

Note: Whenever H_0 is accepted, the capital letter ''Y'' appears in the table with the corresponding confidence level.

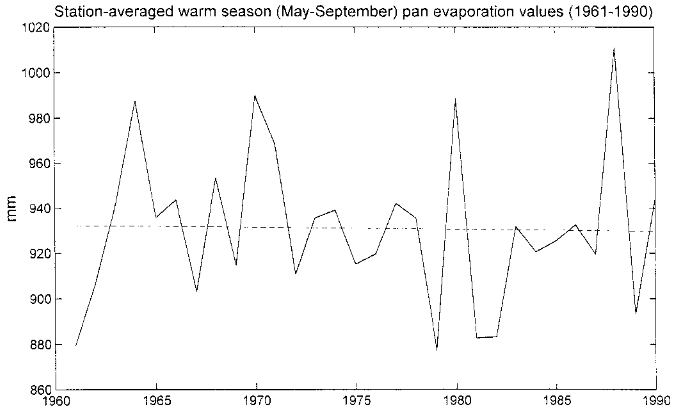

FIG. 9. Warm-Season Pan-Evaporation Values (40 NCDC Stations) and Their Linear Trend Function

areal evapotranspiration fall in line with documented long-term increases in temperature, precipitation, and runoff values, suggesting an accelerated hydrologic cycle (Brutsaert and Parlange 1998) over the conterminous United States.

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NOTATION

The following symbols are used in this paper:

AE = areal evapotranspiration ($M\ L^{-2}\ t^{-1}$);
c_a = heat capacity of air ($E\ M^{-1}\ T^{-1}$);
E_a = drying power of air ($M\ L^{-2}\ t^{-1}$);
E_p = potential evaporation ($M\ L^{-2}\ t^{-1}$);
E_w = wet surface evaporation ($M\ L^{-2}\ t^{-1}$);
G = heat conduction ($E\ L^{-2}\ t^{-1}$);
H = turbulent sensible heat flux ($E\ L^{-2}\ t^{-1}$);
K_E = turbulent transfer coefficient for latent heat ($t^2\ L^{-2}$);
K_H = turbulent transfer coefficient for sensible heat ($E\ T^{-1}\ L^{-3}$);
L = latent heat of vaporization ($E\ M^{-1}$);
P = precipitation ($M\ L^{-2}\ t^{-1}$);
P_a = air pressure ($M\ t^{-2}\ L^{-1}$);
R_n = net energy balance term ($E\ L^{-2}\ t^{-1}$);
R_s = incident global radiation ($E\ L^{-2}\ t^{-1}$);
RO = runoff ($M\ L^{-2}\ t^{-1}$);
TD = temperature difference between surface and air (T);
u = wind velocity ($L\ t^{-1}$);
VPD = vapor pressure deficit ($M\ t^{-2}\ L^{-1}$);
α = Priestley-Taylor constant;
Δ = change in value of variable between adjacent months;
δ = slope of saturation vapor pressure-temperature curve ($M\ t^{-2}\ L^{-1}\ T^{-1}$); and
γ = psychrometric constant ($M\ t^{-2}\ L^{-1}\ T^{-1}$).