Evapotranspiration Intensifies over the Conterminous United States

By Jozsef Szilagyi,[1] Gabriel G. Katul,[2] and Marc B. Parlange[3]

Abstract: Using long-term (1948–1996) pan evaporation measurements, a 6% increase in warm-season (May–October) actual evapotranspiration (*ET*) is computed over the conterminous United States between 1949 and 1996 via the complementary hypothesis. This predicted increase in *ET* is in agreement with the measured precipitation increase for the same period if long-term wet-surface *ET* is assumed to be constant. Long-term relative humidity and air temperature measurements express an increase in mean air temperature and water vapor concentration but not a statistically significant change in vapor pressure deficit. The latter implies a smaller than 6% increase in actual warm-season *ET*. Water-balance estimates for six watersheds, covering about 50% of the land area of the contiguous states of the United States, indicate a 3% increase in annual *ET* over the same period.

INTRODUCTION

Warm-season (May–September) pan evaporation has been documented as decreasing for the past 50 years over parts of North America, and Asia. It has been concluded by Peterson et al. (1995) that such a decline may be caused by an assumed increase in cloudiness (Karl et al. 1993) as a result of ongoing climate change, which leads to a weakened ''terrestrial evaporation component of the hydrologic cycle'' (Peterson et al. 1995). This assertion seemingly contradicts observational data (Karl et al. 1996) and GCM-simulated outputs of increasing precipitation (Manabe 1997).

Recently, Brutsaert and Parlange (1998) resolved the paradox by pointing out that in water-limited environments, pan evaporation is in a complementary relationship with actual evapotranspiration (*ET*), an idea first proposed by Bouchet (1963). Bouchet's hypothesis has repeatedly been shown to accurately predict evaporation (Morton 1983; Parlange and Katul 1992a,b; Kim and Entekhabi 1997). Whether the complementary relationship is indeed consistent with documented climate trends in pan evaporation, precipitation, temperature, water vapor concentration, and water-balance-based basin-scale *ET*, estimates have not been investigated and are the subject of this study. The present study concentrates on the conterminous United States where the decline in pan evaporation values has been the strongest (97 mm per warm season [May–September] in the period 1948–1993 for the western United States and about half of this rate for the eastern part of the United States) among the continental areas investigated by Peterson et al. (1995).

THEORY

Starting with the complementary hypothesis, the actual *ET* (Lt^{-1}) is related to potential (E_p [Lt^{-1}]) evaporation and wet surface evapotranspiration (ET_w [Lt^{-1}]) via

$$ET = 2ET_w - E_p \tag{1}$$

where ET_w can be estimated using a Priestley-Taylor (1972) approach and is given by

$$ET_w = \alpha \frac{\Delta}{\Delta + \gamma} Q_n \tag{2}$$

and E_p is commonly estimated using a Penman-like (1948) combination equation (Katul and Parlange 1992; Parlange and Katul 1992a) or pan evaporation measurements (Katul et al. 1992). If the Penman combination equation is used (instead of pan evaporation), then E_p can be related to mean meteorological conditions using

$$E_p = \frac{\Delta}{\Delta + \gamma} Q_n + \frac{\gamma}{\Delta + \gamma} E_A \tag{3}$$

where E_A is the drying power of the atmosphere and can be expressed as (Brutsaert 1982; Katul and Parlange 1992)

$$E_A = f_e(u)(e^* - e) = f_e(u)(VPD) \tag{4}$$

with $f_e(u)$ [L^2tM^{-1}] defined for a neutral atmosphere as

$$f_e(u) = \frac{0.622 a_v k^2 u}{\rho_w R_d T_a \ln\left(\frac{z - d_o}{z_{om}}\right) \ln\left(\frac{z - d_{ov}}{z_{ov}}\right)} \tag{5}$$

where u (Lt^{-1}) = the mean horizontal wind speed; a_v ($\approx 1[-]$) = the inverse of the turbulent Schmidt number under neutral conditions; k ($=0.4[-]$) = the Von Karman constant; ρ_w (ML^{-3}) = water density; R_d ($ET^{-1}M^{-1}$) = the gas constant for dry air; T_a = the mean air temperature; e^* and e = the saturation and actual vapor pressure at T_a; z = height above the surface, d_o and d_{ov} = the zero-plane displacements (L) for momentum and water vapor, respectively; z_{ov} and z_{om} = the water vapor and momentum roughness lengths (L); α ($\approx 1.26[-]$) is the Priestley-Taylor constant (Katul and Parlange 1992; Parlange and Katul 1992a,b); Δ (PT^{-1}) = the slope of the saturation vapor pressure-temperature curve; γ (PT^{-1}) = the psychrometric constant; *VPD* (P) is the vapor pressure deficit; and Q_n is the available energy. The near-neutral approximation in (5) is reasonable for long averaging times given the close proximity of the meteorological measurements to the ground surface and hence, the more important role of mechanical production of turbulent kinetic energy when compared to buoyant production. Upon replacing (2)–(4) in (1), a relationship between actual *ET* and mean meteorological conditions can be derived and is given by

$$ET = \left(2\alpha \frac{\Delta}{\Delta + \gamma} - 1\right) Q_n - \frac{\gamma}{\Delta + \gamma} E_A = 2ET_w - E_p \tag{6}$$

[1]Conservation and Survey Div., Univ. of Nebraska, Lincoln, NE 68588-0517. E-mail: jszilagyil@unl.edu

[2]School of the Envir., Duke Univ., Durham, NC 27708. E-mail: gaby@duke.edu

[3]Dept. of Geography and Envir. Engrg., The Johns Hopkins Univ., Baltimore, MD 21218-2686. E-mail: mbparlange@jhu.edu

Note. Discussion open until May 1, 2002. To extend the closing date one month, a written request must be filed with the ASCE Manager of Journals. The manuscript for this paper was submitted for review and possible publication on May 22, 2000; revised November 14, 2000. This paper is part of the ***Journal of Water Resources Planning and Management***, Vol. 127, No. 6, November/December, 2001. ©ASCE, ISSN 0733-9496/01/0006-0354–0362/$8.00 + $.50 per page. Paper No. 22270.

A common feature of the National Climatic Data Center (NCDC) data set is that measurements are missing between 1965 and 1983. This results in 16 years of continuous data prior to 1965 and in 13 years of continuous data in the period 1984–1996 for analysis. For this reason, relative changes in computed *ET* for these two periods (1949–1964 and 1984–1996) are compared with the measured relative change in precipitation between the same periods, as discussed next.

APPLICATION OF THE COMPLEMENTARY RELATION

Long-term temporal changes in *ET* ($=d\langle\overline{ET}\rangle/dt$ can be quantified from (6) using

$$\frac{d\langle\overline{ET}\rangle}{dt} = \frac{\partial ET}{\partial Q_n}\frac{d\langle\overline{Q_n}\rangle}{dt} + \frac{\partial ET}{\partial E_A}\frac{d\langle\overline{E_A}\rangle}{dt} \tag{7a}$$

$$\frac{d\langle\overline{ET}\rangle}{dt} = 2\,\frac{d\langle\overline{ET_w}\rangle}{dt} - \frac{d\langle\overline{E_p}\rangle}{dt} \tag{7b}$$

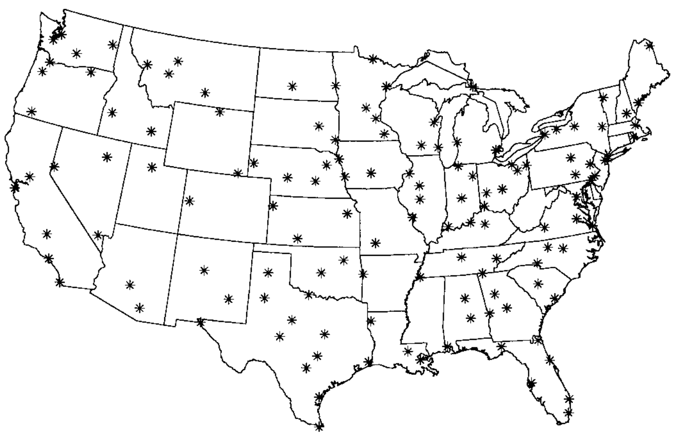

FIG. 1. Locations of the 132 Meteorological Stations Employed in the Study

where the over-bar implies temporal averaging over a long period (>10 years) and $\langle\cdot\rangle$ implies spatial averaging. An implicit assumption in the Brutsaert and Parlange (1998) argument is that ET_w does not vary appreciably between the two periods ($d\langle\overline{ET_w}\rangle/dt = 0$) so that changes in *ET* are primarily due to changes in potential evaporation as evidenced by (7*a*). Perusing the records of more than 500 meteorological stations across the conterminous United States, we found 132 stations (Fig. 1) that had long-term measurements, generally starting in 1949, of daily maximum and minimum temperature, as well as maximum and minimum relative humidity and precipitation. The NCDC data set contains only 40 stations over the conterminous United States with long-term pan evaporation data. We found this number too low to use the NCDC pan evaporation data as representative for the conterminous United States. Instead, we relied upon the pan data that was published by Peterson et al. (1995).

Using a 6% decrease in long-term, warm-season pan evaporation measurements from Peterson et al. (1995), and assuming $d\langle\overline{ET_w}\rangle/dt = 0$ in (7*b*), the observed reduction in E_p corresponds to a symmetric increase of 6% in ET (provided it is further assumed that pan evaporation is a ''perfect'' estimator of E_p). This seems to be in agreement with the 11% measured increase in precipitation between the two periods (Fig. 2, Table 1), because not all the precipitation would be recycled as *ET*. Hence, it appears that the complementary relationship in (1) does, in a quantitative manner, resolve the hydrologic paradox if $d\langle\overline{ET_w}\rangle/dt = 0$. Next we investigate the consequence of a constant ET_w in the context of (7*a*).

DISCUSSION

From (2), an immediate consequence of $d\langle\overline{ET_w}\rangle/dt = 0$ is that $d\langle\overline{Q_n}\rangle/dt = 0$, if the change in $\Delta/(\Delta + \gamma)$ is neglected between the two periods. Hence, from (7*a*), $d\langle\overline{ET}\rangle/dt$ is primarily driven by $d\langle\overline{E_A}\rangle/dt$, or stated differently

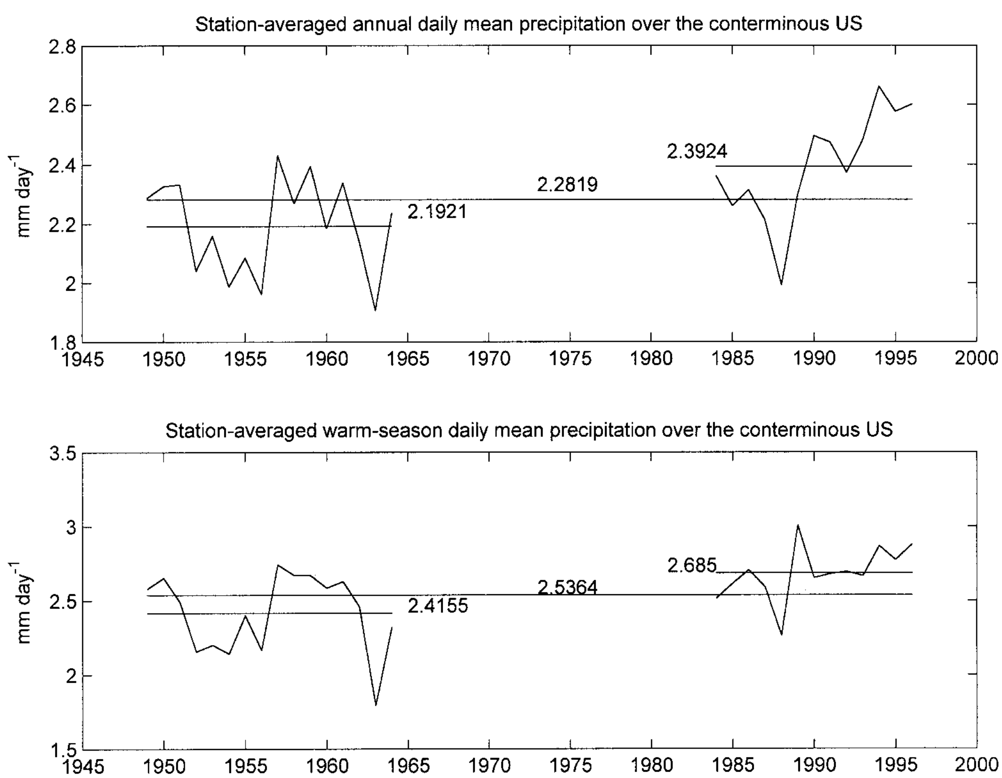

FIG. 2. Station-Averaged Annual and Warm-Season (May–September) Daily Mean Precipitation Time Series for the Conterminous United States

TABLE 1. Station-Averaged (N = 132) Hydro-Meteorological Variables, Their Relative Changes between the Periods, and the Accompanying Two Sample t-Test Results at 95% Significance Level

Hydro-meteorological variables	1949–1964	1984–1996	Absolute and relative (%) changes (=$[V_2 - V_1]/V_1$) between periods	t-test results: H_0: $V_2 \neq V_1$
Humidity (HPa)				
Annual	11.67	12.02	+0.35, +3.00	Yes
May–September	17.06	17.69	+0.63, +3.69	Yes
Temperature (F)				
Annual	55.24	55.76	+0.52, +0.94	No
May–September	70.52	70.90	+0.38, +0.54	No.
Vapor pressure deficit (HPa)				
Annual	6.03	5.97	−0.06, −1.00	No
May–September	9.50	9.24	−0.26, −2.74	No
Daily mean precipitation (mm day^{-1})				
Annual	2.19	2.39	+0.20, +9.13	Yes
May–September	2.42	2.69	+0.27, +11.16	Yes

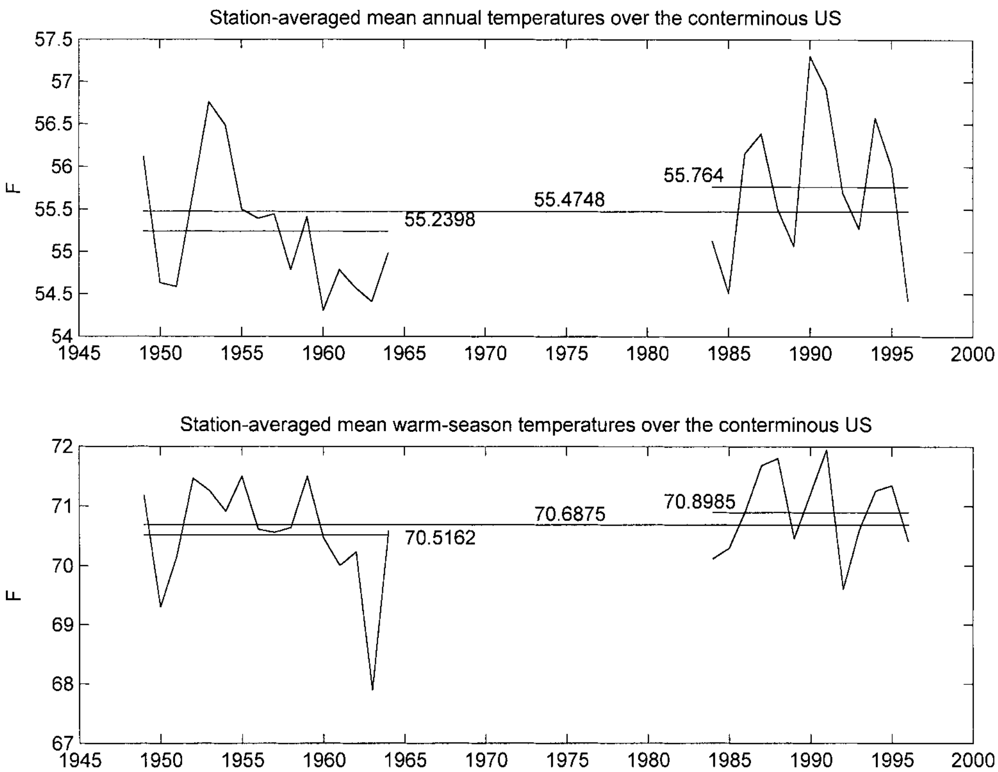

FIG. 3. Station-Averaged Mean Annual and Warm-Season (May–September) Air Temperature Time Series for the Conterminous United States

$$\frac{d\langle \overline{ET} \rangle}{dt} = -\frac{\gamma}{\Delta + \gamma}\left(\langle \overline{f_e(u)} \rangle \frac{d\langle \overline{VPD} \rangle}{dt} + \langle \overline{VPD} \rangle \frac{d\langle \overline{f_e(u)} \rangle}{dt}\right) \quad (8)$$

assuming that: (a) changes in the wind function are decoupled from changes in the vapor pressure deficit; and (b) the change in the $\gamma/(\Delta + \gamma) = 1 - \Delta/(\Delta + \gamma)$ term is negligible between the two periods. In a first order analysis, it is here assumed that changes in wind patterns are less significant than changes in mean air temperature and water vapor concentration so that the second term on the right-hand side of (8) is negligible relative to the first term. To assess how much of the change in evaporation is due to changes in *VPD*, daily mean vapor pressure, e (KPa), was calculated using

$$e = \hat{r}a10^{b\hat{T}/c+\hat{T}} = \hat{r}e^* \quad (9)$$

where $r(-)$ = the relative humidity; a, b, and c = empirical constants $(-)$ with values of 0.61, 7.5 and 237.3, respectively (Tetens 1930); T = temperature in Celsius; and e^* is the corresponding saturation vapor pressure. The had in the case of the relative humidity and temperature data denotes the arithmetic averages of the daily maximum and minimum values recorded.

Long-term increases in air temperature (Fig. 3, Table 1) and water vapor concentration (Fig. 4, Table 1) were both quantified over the conterminous United States with computed changes in both variables and are comparable to other published results (Gaffen and Ross 1998, 1999). However, despite the statistically significant increase in water vapor concentration, we found that *VPD* stayed practically constant between these two periods (Fig. 5, Table 1), especially on an annual basis. A direct consequence of the near constant *VPD* ($d\langle \overline{VPD} \rangle/dt \approx 0$) is that $d\langle \overline{E_A} \rangle/dt \approx 0$ (again assuming the wind pattern is identical for these two averaging periods). Unfortunately, wind speed measurements were not available for the first period (1949–1964) to directly test the validity of this conjecture.

The broader implication of our finding is that despite the good agreement between predicted increase in ET using the complementary relationship and measured increase in precipitation, the available energy Q_n might have easily changed between the two periods. A change in ET_w resulting from a change in Q_n cannot be consistent with the 6% increase in ET predicted by the complementary relationship. In fact, a slight reduction (provided $2|d\langle \overline{ET_w} \rangle/dt| < |d\langle \overline{E_p} \rangle/dt|$) in ET_w causes a

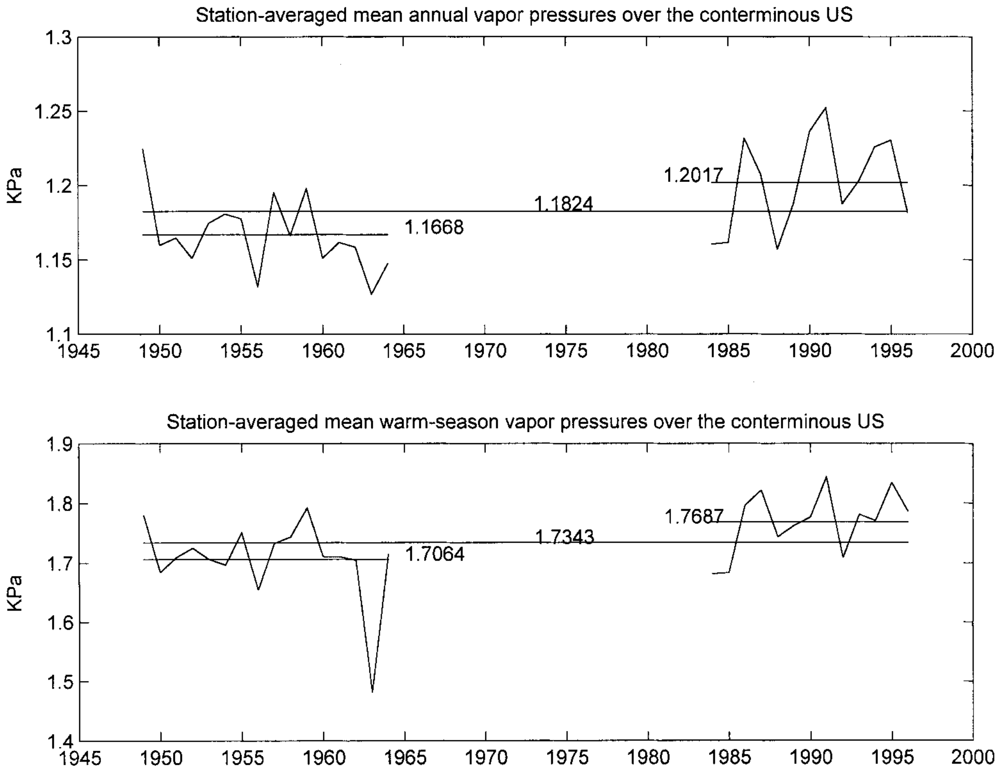

FIG. 4. Station-Averaged Mean Annual and Warm-Season (May–September) Vapor Pressure Time Series for the Conterminous United States

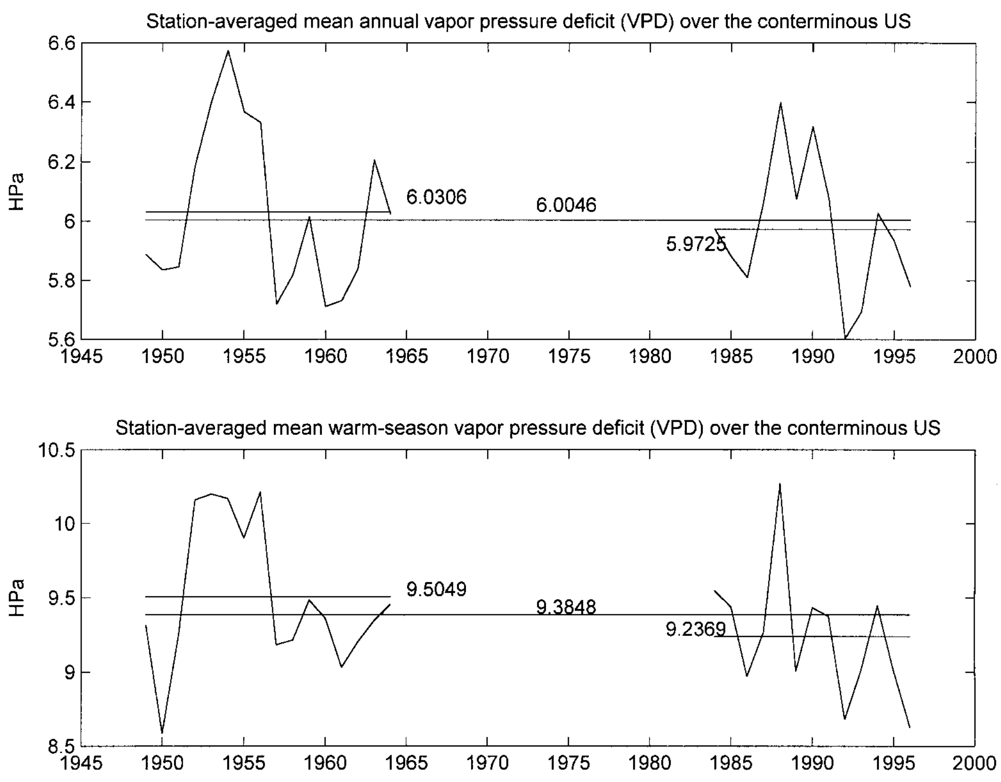

FIG. 5. Station-Averaged Mean Annual and Warm-Season (May–September) Vapor Pressure Deficit Time Series for the Conterminous United States

diminished increase in actual *ET* due to (6). Such a Q_n decrease is consistent with a possible, but yet unsubstantiated, increase in cloudiness (Karl et al. 1993; Peterson et al. 1995). In conclusion, the complementary relationship predicted of increased *ET* using pan evaporation data may overestimate the real increase in *ET* over the conterminous United States when a constant Q_n term is used. This conclusion is corroborated by the results of water-balance calculations for six medium-to-large size watersheds within the conterminous United States.

Water-year (October–September) *ET* for each watershed was calculated as the difference between precipitation and streamflow. In the absence of any reported continental-scale long-term trends in groundwater-table elevations, changes in stored water volumes were considered to be negligible among consecutive water-years.

Care had to be taken with the selection of watersheds. Ideally, the larger the watershed the larger the area about which one can obtain information on *ET*; therefore, by concentrating

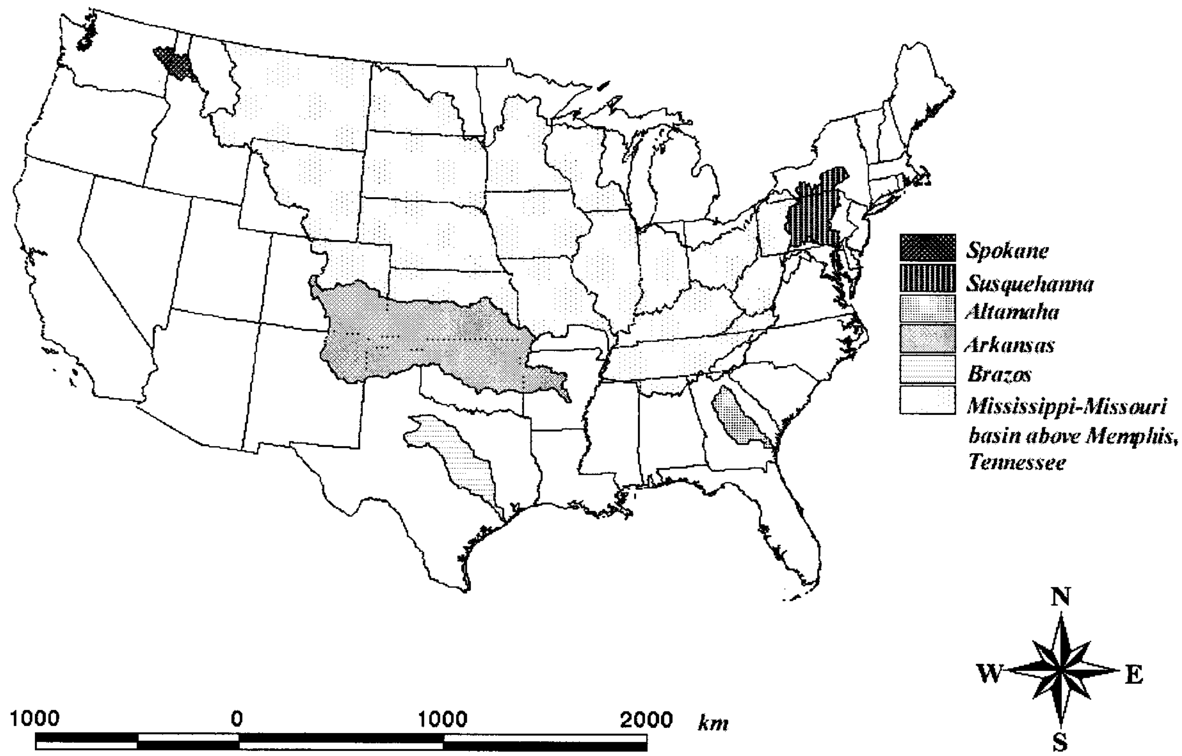

FIG. 6. Locations of Watersheds Studied

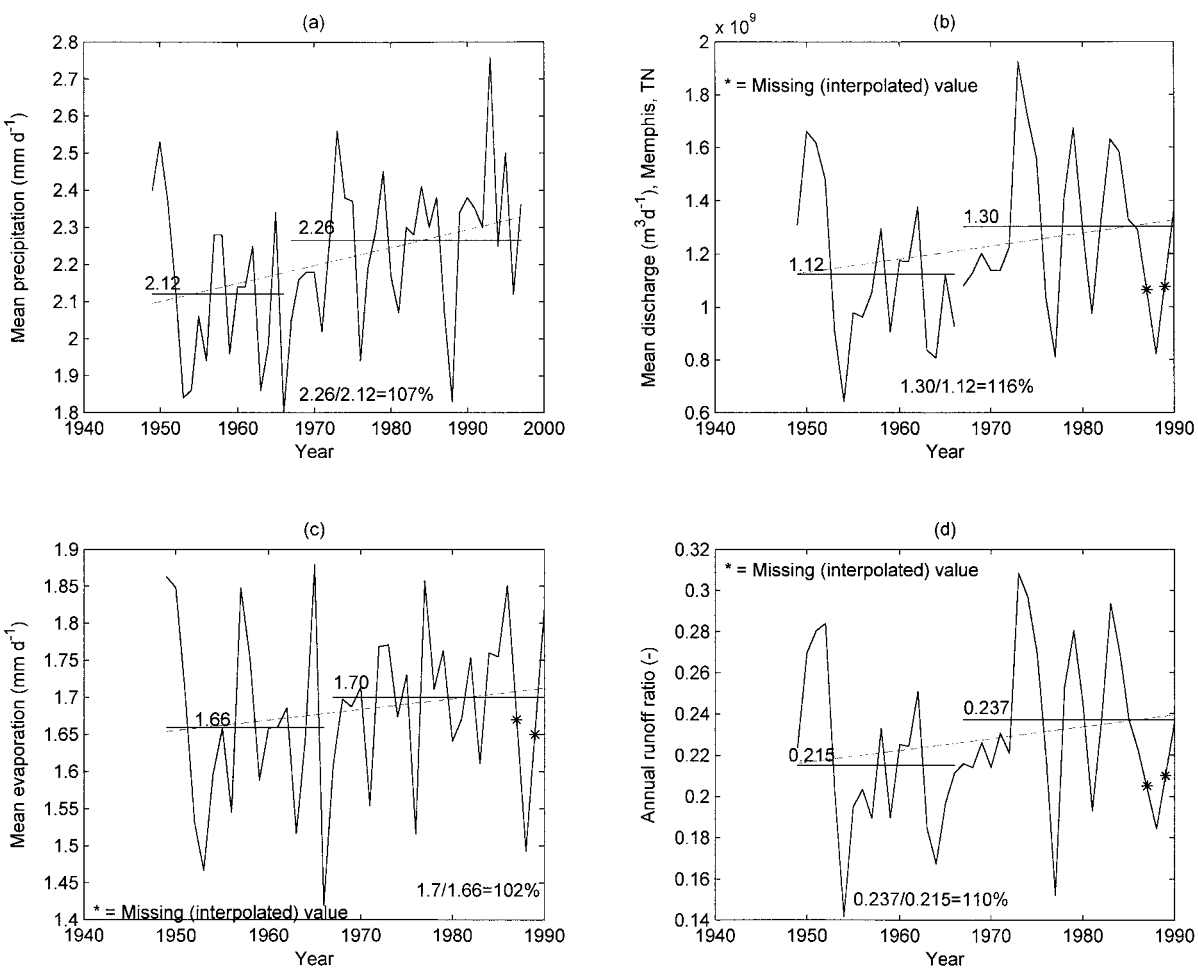

FIG. 7. Mississippi-Missouri River Basin above Memphis, Tennessee

on the largest catchments, one could cover almost the entire conterminous United States through the selection of only a few large catchments, such as the Mississippi-Missouri basin. In practice, however, watersheds, especially in the semiarid Southwest, are so disturbed by human intervention that one may become sceptical about the validity of any water-balance calculations for the present purpose. Consider, for example, the Colorado River (the third largest river in the conterminous United States) basin where the river reaches its original destination, the Gulf of California, only in years with unusually high precipitation (Reisner 1993).

Fig. 6 displays the geographic location of the six watersheds within the contiguous United States where years with incomplete data in either the daily precipitation or stream discharge variables made up less than 5% (3 years) of the period 1948–1997. Table 2 lists the location of the gauging stations, the corresponding drainage areas, as well as the number of precipitation stations used with each catchment to calculate annual daily mean evapotranspiration. The six watersheds selected cover about 50% of the total land area of the conterminous United States. Unfortunately, the second- to fourth-largest basins (the Columbia, Colorado, and Rio Grande) had to be omitted from the analysis. The last two were omitted for reasons detailed above, the first due to the lack of long-term discharge data. Many other large watersheds had to be excluded from the study as well, due to data insufficiency, such as (counterclockwise around the Mississippi-Missouri basin) the Willamette, Sacramento, Red, Mobile-Alabama, James-Jackson, Hudson, and Connecticut rivers, etc. Even for the Mississippi-Missouri basin, the lowest downstream gauging station with 50 years of discharge data was as far upstream as Memphis, Tennessee.

Figs. 7 through 12 display the daily mean precipitation, discharge, estimated ET, annual runoff ratios (stream discharge divided by precipitation), as well as their means over two subperiods, 1949–1966 and 1967–1997, respectively. These data were also accompanied by a linear trend function in each case for the six watersheds.

As can be seen, precipitation has increased over the past 50 years in all watersheds, except the Spokane River basin. Similarly, stream discharge and the runoff ratio, as a consequence of enhanced precipitation, increased in all watersheds except

TABLE 2. Gauging Station Locations, Contributing Drainage Area, and Number of Precipitation Stations within the Watershed

Gauging station location	Contributing drainage area (km^2)	Number of precipitation stations in the watershed
Altamaha River basin near Doctortown, GA	35,224	2
Arkansas River basin near Little Rock, AK	351,690	5
Brazos River basin near Richmond, TX	91,792	3
Mississippi-Missouri River basin at Memphis, TN	2,437,200	33
Spokane River basin at Long Lake, WA	15,592	1
Susquehanna River basin at Harrisburg, PA	62,419	2

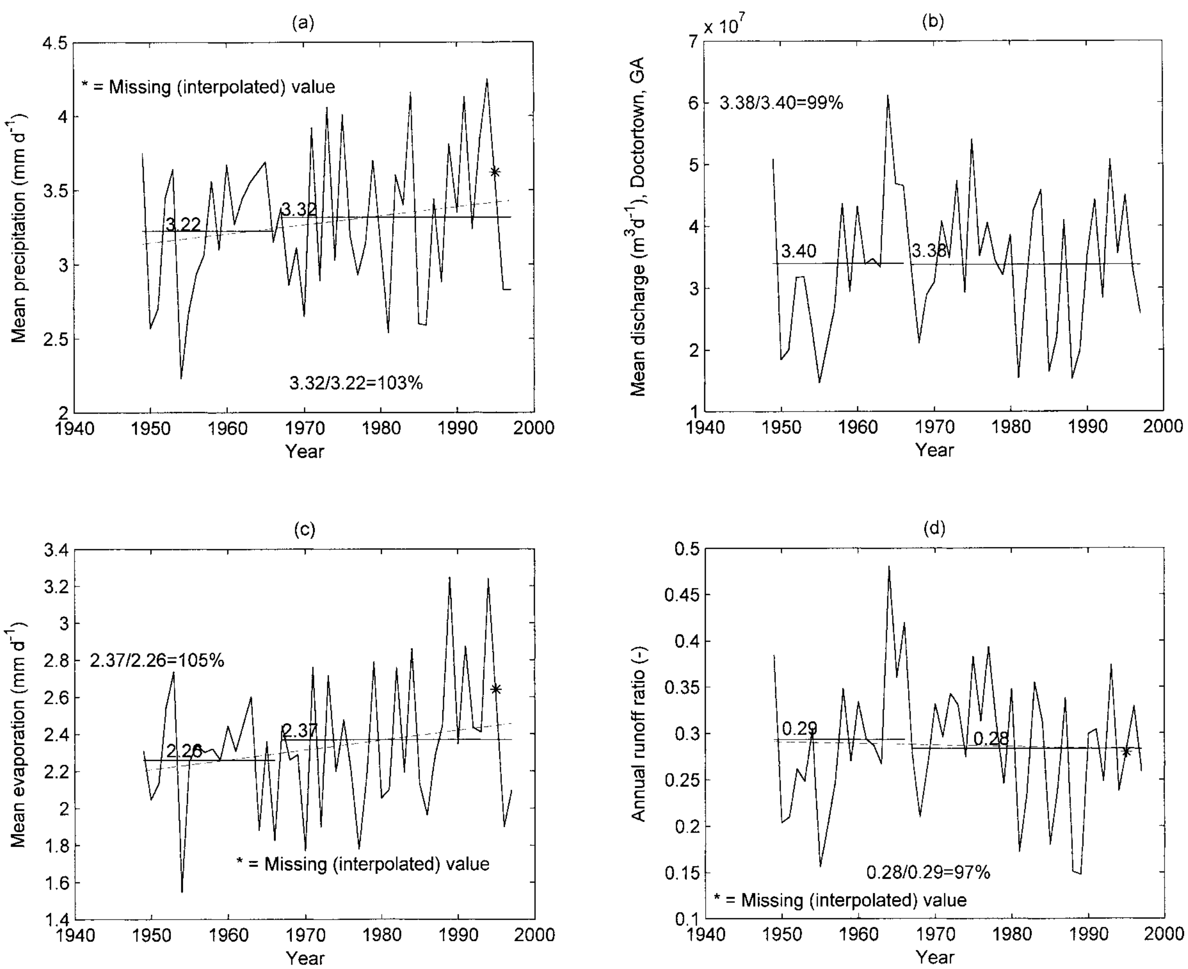

FIG. 8. Altamaha River Basin above Doctortown, Georgia

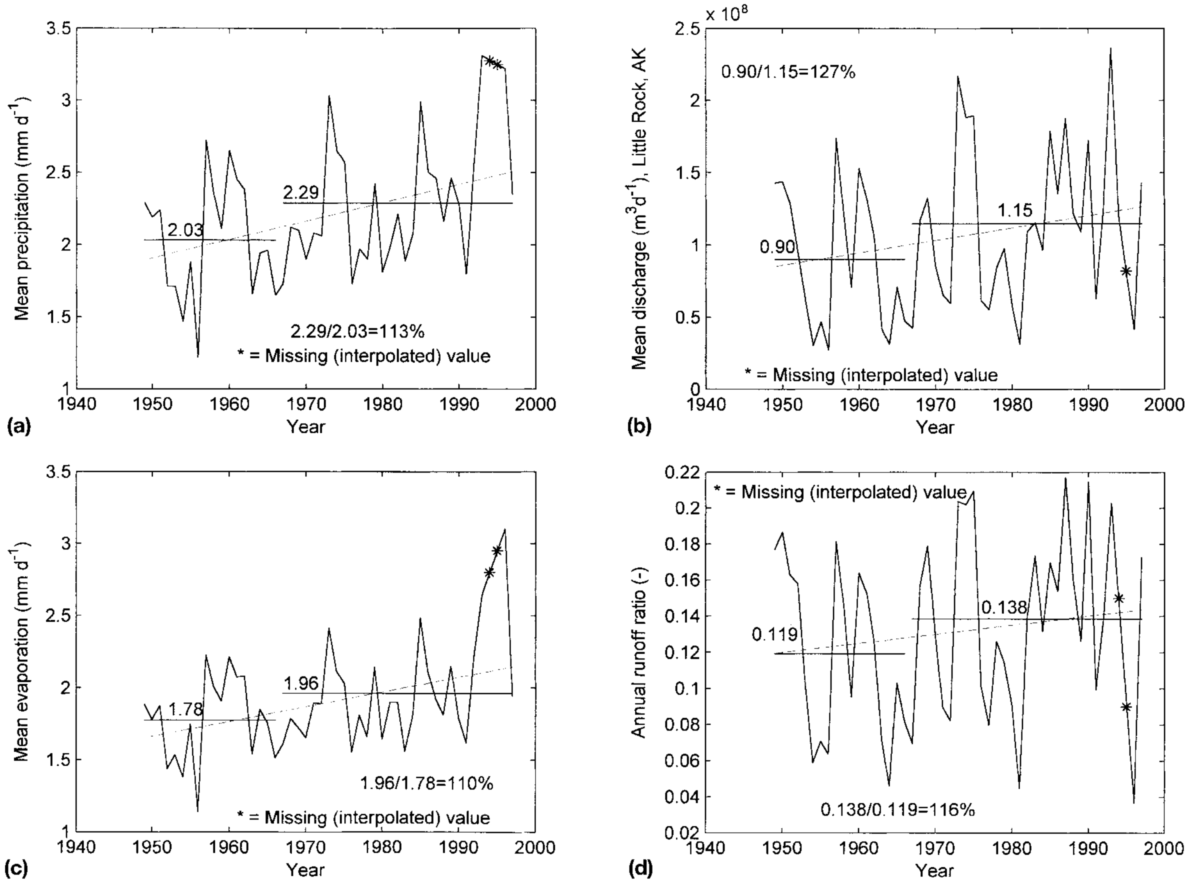

FIG. 9. Arkansas River Basin above Little Rock, Arkansas

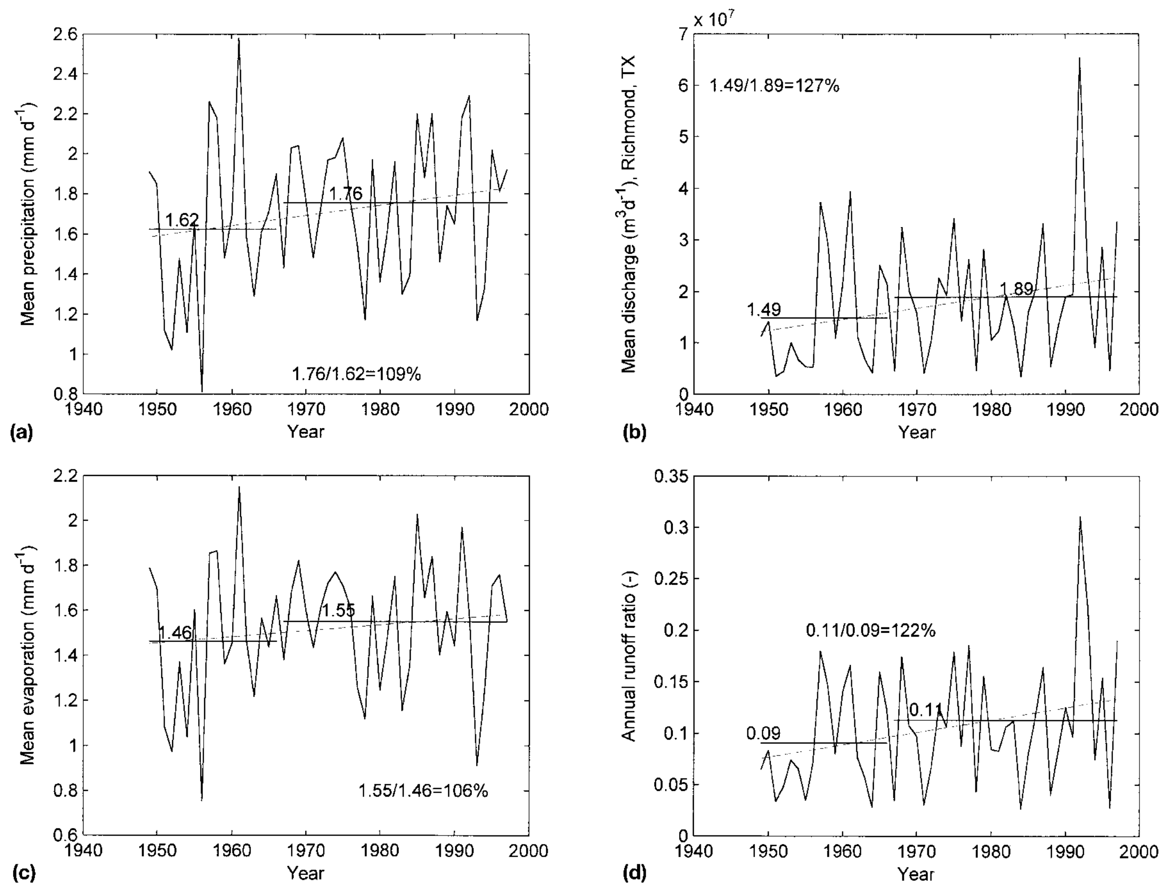

FIG. 10. Brazos River Basin above Richmond, Texas

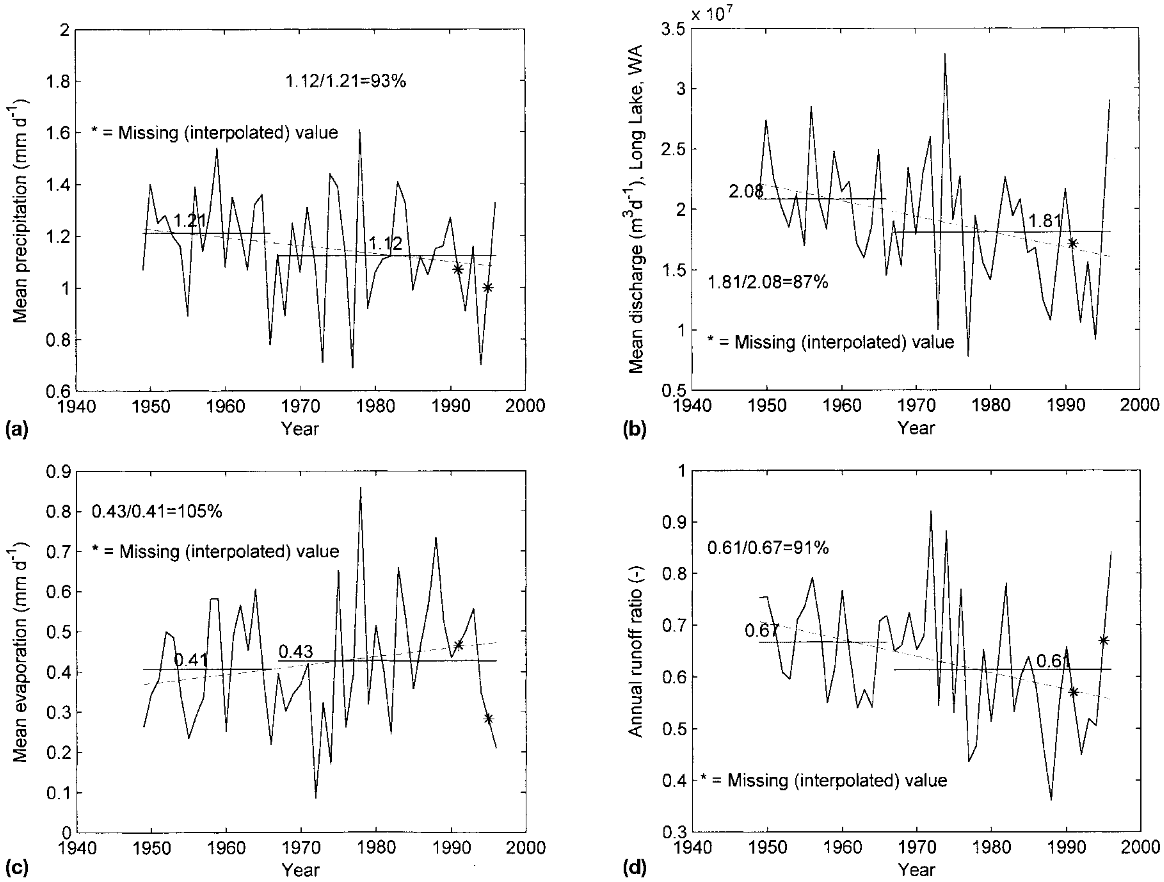

FIG. 11. Spokane River Basin above Long Lake, Washington

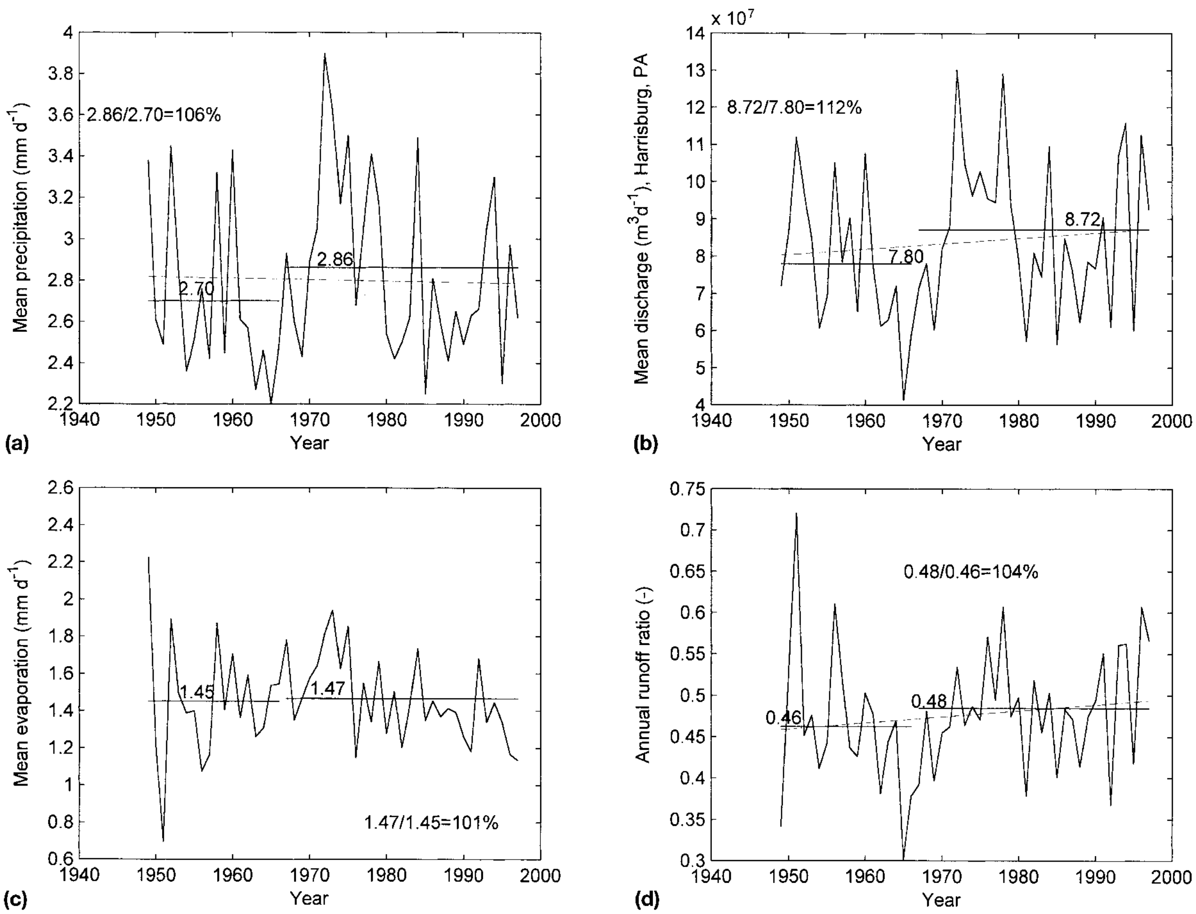

FIG. 12. Susquehanna River Basin above Harrisburg, Pennsylvania

TABLE 3. Results of the Two Sample t-Tests; Daily Mean Water-Year Evaporation (mm)

Basin	Mean (1949–1966) (m_1)	Mean (1967–1997) (m_2)	m_2/m_1 (%)	Significance level (%) that $m_1 \neq m_2$
Altamaha	2.26	2.37	105	69
Arkansas	1.78	1.96	110	93
Brazos	1.46	1.55	106	67
Mississippi-Missouri	1.66	1.70	102	61
Spokane	0.41	0.43	105	33
Susquehanna	1.45	1.47	101	14
Area-weighted average	—	—	103	—

in the Spokane River basin, where they declined, and in the Altamaha River basin, where they stayed practically constant. These results agree with Lettenmaier et al. (1994), who studied trends in the two variables for 1948–1988 within the same geographic area. Long-term *ET* displays a unanimously increasing trend in all watersheds studied. Table 3 summarizes the changes. While a ubiquitous increase (with a mean relative change of 5%) in the subperiod-averaged values is clearly visible in the graphs, the corresponding two sample *t*-test significance levels of the assumed change in the mean values, with the exception of the Arkansas River basin, are generally low.

SUMMARY

Application of Bouchet's (1963) complementary hypothesis, assuming a constant net energy term, with long-term hydrometeorological data predicts a 6% increase in warm-season (May–October) *ET* over the conterminous United States over approximately the past 50 years. An increased level of *ET* is in accordance with observed increases in precipitation and temperature levels (Karl et al. 1996) over the same area. A near constant mean long-term vapor pressure deficit, however, suggests that a constant mean net radiation balance may not be assumed over the same period, implying that the complementary hypothesis probably overestimates the increase in long-term *ET*. This conclusion seems to be corroborated by the result of water-balance-derived basin-scale *ET* calculations for six watersheds within the conterminous United States covering about 50% of the total land area. Area-weighted mean long-term watershed *ET* shows an overall 3% increase in annual watershed *ET* over the past 50 years. This observed 3% average increase in *ET*, however, is laden with great uncertainty due to the large range in the year-to-year variability of the estimated *ET* values. At the same time, an increase in calculated watershed *ET* is present at *all* watersheds studied.

Based on the analysis presented here, it is concluded that an increase in long-term *ET* is detectable over the conterminous United States. The precise rate of that increase, however, remains uncertain due to the generally large level of interannual variation in the areal evapotranspiration estimates.

ACKNOWLEDGMENT

One of the authors is grateful to Charles Flowerday for his editorial comments on the manuscript.

REFERENCES

Bouchet, R. J. (1963). ''Evapotranspiration reelle at potentielle, signification climatique.'' *Int. Assoc. Sci. Hydro. Pub.*, 62, 134–142.

Brutsaert, W., and Parlange, M. B. (1998). ''Hydrologic cycle explains the evaporation paradox.'' *Nature*, 396, 30.

Gaffen, D. J., and Ross, R. J. (1999). ''Climatology and trends of US surface humidity and temperature.'' *J. Climate*, 12, 811–828, 1999.

Gaffen, D. J., and Ross, R. J. (1998). ''Increased summertime heat stress in the US.'' *Nature*, 396, 529.

Karl, T. R., Jones, P. D., Knight, R. W., and Kukla, G. (1993). ''A new perspective on recent global warming—asymmetric trends of daily maximum and minimum temperatures.'' *Bull. Am. Meteorol. Soc.*, 74, 1007–1023.

Karl, T. R., Knight, R. W., Easterling, D. R., and Quayle, R. G. (1996). ''Indices of climate change for the United States.'' *Bull. Am. Meteorol. Soc.*, 77, 279–292.

Katul, G. G., and Parlange, M. B. (1992). ''A Penman-Brutsaert model for wet surface evaporation.'' *Water Resour. Res.*, 28, 121–126.

Katul, G. G., Cuenca, R. H., Grebet, P., Wright, J. L., and Pruitt, W. O. (1992). ''Analysis of evaporative flux data for various climates.'' *J. Irrig. and Drain. Engrg.*, ASCE, 118(4), 601–618.

Kim, C. P., and Entekhabi, D. (1997). ''Examination of two methods for estimating regional evaporation using a coupled mixed layer and land surface model.'' *Water Resour. Res.*, 33, 2109–2116.

Lettenmaier, D. P., Wood, E. F., and Wallis, J. R. (1994). ''Hydro-climatological trends in the continental United States, 1948–88.'' *J. Climate*, 7, 586–607.

Manabe, S. (1997). ''Study of global warming by GFDL climate models.'' *Ambio*, 26, 47–51.

Morton, F. I. (1983). ''Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology.'' *J. Hydrol.*, Amsterdam, 66, 1–76.

Parlange, M. B., and Katul, G. G. (1992a). ''An advection-aridity evaporation model.'' *Water Resour. Res.*, 28, 127–132.

Parlange, M. B., and Katul, G. G. (1992b). ''Estimation of the diurnal variation of potential evaporation from a wet bare soil surface.'' *J. Hydrol.*, 132, 71–89.

Penman, H. L. (1948). ''Natural evaporation from open water, bare soil, and grass.'' *Proc., Roy. Soc. London*, A193, 120–146.

Peterson, T. C., Golubev, V. S., and Groisman, P. Y. (1995). ''Evaporation losing its strength.'' *Nature*, 377, 687–688.

Priestley, C. H. B., and Taylor, R. J. (1975). ''On the assessment of surface heat flux and evaporation using large-scale parameters.'' *Month. Weather Rev.*, 100, 81–92.

Reisner, M. (1993). *Cadillac Desert*, Penguin Books, New York.

Tenes, O. (1930). ''Uber einige meteorologische Begriffe.'' *Z. Geophys.*, 6, 203–204.

NOTATION

The following symbols are used in this paper:

a, b, c = empirical constants [−];
a_v = inverse of the turbulent Schmidt number [−];
d_o, d_{ov} = zero-plane displacements for momentum and water vapor, respectively [L];
E_p = potential evaporation [Lt^{-1}];
ET = actual evapotranspiration [Lt^{-1}];
ET_w = wet surface evapotranspiration [Lt^{-1}];
e^*, e = saturation and actual vapor pressure, respectively [P];
k = Von Karman constant [−];
Q_n = net energy [E];
R_d = gas constant for dry air [ET^{-1} M^{-1}];
r = relative humidity [−];
T_a = mean air temperature [T];
VPD = vapor pressure deficit [P];
t = time [t];
z = height above the surface [L];
z_{ov}, z_{om} = water vapor and momentum roughness lengths, respectively [L];
α = Priestley-Taylor constant [−];
γ = psychrometric constant [PT^{-1}]; and
Δ = slope of the saturation vapor pressure-temperature curve [PT^{-1}].