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Hydroclimatology of the continental United States

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[1] The overall water balance and the sensitivity of watershed runoff to changes in climate are investigated using national databases of climate and streamflow for 1,337 watersheds in the U.S. We document that 1% changes in precipitation result in 1.5–2.5% changes in watershed runoff, depending upon the degree of buffering by storage processes and other factors. Unlike previous research, our approach to estimating climate sensitivity of streamflow is nonparametric and does not depend on a hydrologic model. The upper bound for precipitation elasticity of streamflow is shown to be the inverse of the runoff ratio. For over a century, investigators [Pike, 1964; Budyko, 1974; Ol'dekop, 1911; and Schreiber, 1904] have suggested that variations in watershed aridity alone are sufficient to predict spatial variations in long-term watershed runoff. We document that variations in soil moisture holding capacity are just as important as variations in watershed aridity in explaining the mean and variance of annual watershed runoff.

INDEX TERMS: 1833 Hydrology: Hydroclimatology; 1836 Hydrology: Hydrologic budget (1655); 1860 Hydrology: Runoff and streamflow; 1878 Hydrology: Water/energy interactions; **KEYWORDS:** hydroclimatology, hydrologic budget, water/energy interactions, runoff and streamflow.
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1. Introduction

[2] Our understanding of hydroclimatic processes over broad spatial scales is important within the context of: global warming [Wigley and Jones, 1985]; efforts to understand the continental and global water and energy cycles [National Research Council, 1998]; and the increasing demand for water due to population and economic growth [Watson et al., 2000]. Our understanding of hydrological processes at continental scales has been constrained due to our inadequate understanding of hydrologic processes and due to difficulties in estimating hydrologic fluxes [Roads et al., 1994]. Recent advances in hydrologic data collection using remote sensing techniques and concerted efforts by national and international agencies together with the development of improved techniques for converting point estimates of hydrological fluxes to spatial estimates provide an opportunity to understand hydroclimatic issues particularly at continental and global scales. We present an overview of the hydroclimatology of the continental United States and summarize recent develop-

ments in modeling the long-term water balance, the inter-annual variability of streamflow and the sensitivity of watershed runoff to changes in precipitation over the U.S.

2. National Hydroclimatic Database

[3] To support this research, high quality time series of monthly streamflow, precipitation and temperature were constructed to account for the complex spatial and temporal variations in hydroclimatology of the continental U.S. Daily streamflow records were obtained from a national Hydro-Climatic Data Network (HCDN) for 1,337 basins in the continental U.S. [Slack et al., 1993]. This database is unique because it is relatively free from anthropogenic influences including: ground water pumping, flow diversion, and/or land use changes. For each watershed, 37-year monthly time series of precipitation, average maximum daily temperature and average minimum daily temperature were derived from 0.5-degree time-series grids based on the PRISM climate modeling system [Daly et al., 1994]. The monthly climatic time series grids were spatially averaged over each HCDN watershed. The watershed boundaries were delineated using a 1 km. digital elevation map of the U.S. For more information regarding the development of the hydroclimatic database and/or the location of the 1,337 watersheds, see [Sankarasubramanian and Vogel, 2002]. Using monthly time series of average minimum and average maximum temperature data along with extraterrestrial solar radiation, monthly time series of potential evapotranspiration were obtained using the method introduced by Hargreaves and Samani [1982]. The result is a unique set of time-series of monthly precipitation, potential evapotranspiration and streamflow over the period 1951–1988 for 1,337 watersheds distributed across the U.S.

3. Climate Elasticity of Streamflow

[4] Until reliable climate model output is available, the sensitivity of streamflow to changes in climate is perhaps best understood using the historical records of streamflow and climate [Risbey and Entekhabi, 1996]. Climate elasticity of streamflow, an index commonly used to quantify the sensitivity of streamflow to changes in climate, may be defined as the proportional change in streamflow, Q , to the proportional change in a climatic variable such as precipitation P in (1).

$$\varepsilon_P(P, Q) = (\partial Q / \partial P) \cdot (P / Q) \quad (1)$$

[5] A problem in estimating climate elasticity using (1) is that the differential is model dependent, and of course, the structure of the hydrologic model is always unknown. Recent research documents that the non-parametric estimator

$$e_P = \text{median}[(Q_t - \bar{Q}) / (P_t - \bar{P}) \cdot (\bar{Q} / \bar{P})] \quad (2)$$

is robust for estimating the climate elasticity of streamflow for a wide class of hydrologic models [Sankarasubramanian et al., 2001].

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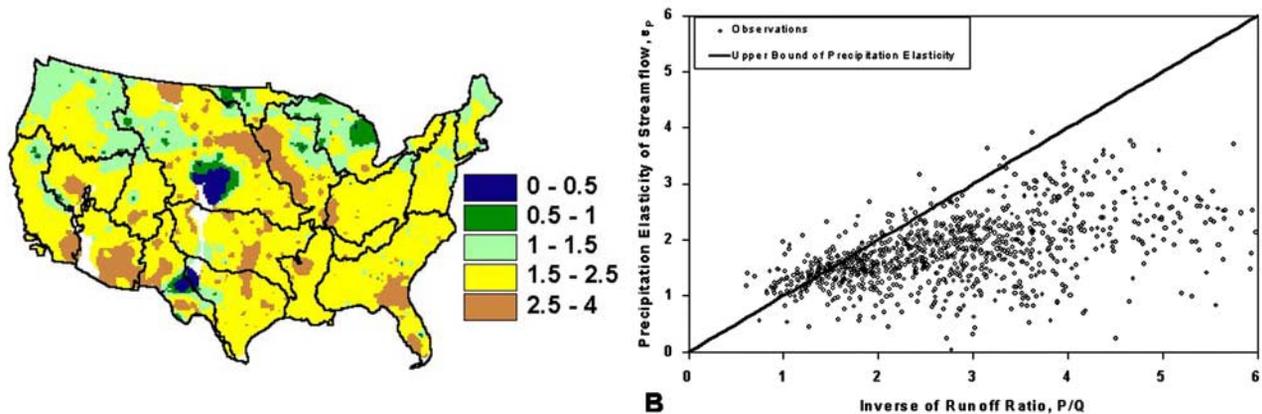


Figure 1. Sensitivity of streamflow to changes in precipitation in the continental U.S. (a) Precipitation elasticity of streamflow (b) Upper bound for precipitation elasticity of streamflow.

[6] Estimates of the precipitation elasticity of streamflow $\varepsilon_P(\mu_P, \mu_Q)$ for the 1,337 HCDN river basins are obtained using (2) and the results are illustrated as a map in Figure 1a for the continental United States. A value of $\varepsilon_P(\mu_P, \mu_Q)$ greater than 1 indicates that a 1% change in precipitation will result in more than a 1% change in streamflow. Values of $\varepsilon_P(\mu_P, \mu_Q)$ generally range from 1.0–2.5 across the U.S., implying that the precipitation-runoff relationship is generally non-linear with the nonlinearity influenced by storage processes within the basin. For example, storage and soil properties act to buffer the impacts of climate change so that e_P is lower in the Sand Hills region of Nebraska. Arid and semi-arid basins in the midwestern and southwestern regions of the U.S. exhibit greater precipitation elasticities than the humid eastern and northwestern regions. Comparisons with 10 previous independent climate change studies revealed that Figure 1a can also provide a useful validation metric for past and future water resource related climate change investigations in the U.S. [Sankarasubramanian *et al.*, 2001].

[7] Figure 1b illustrates that an upper bound for precipitation elasticity is the reciprocal of the runoff ratio. Suppose interannual variations in watershed storage processes are negligible compared to interannual variations in precipitation. For a very humid basin with moisture available throughout the year, the maximum incremental streamflow resulting from a 1% increase in precipitation is $0.01P$. This will produce a corresponding increase in runoff, $\delta Q/Q$, equal to $0.01P/Q$, and by substituting this quantity into (1), we obtain an upper bound for the precipitation elasticity of streamflow as the inverse of the runoff ratio or \bar{P}/\bar{Q} . With the exception of only a few basins in northwestern U.S., Figure 1b illustrates that for most basins in the U.S., the estimator e_P falls below the upper bound \bar{P}/\bar{Q} . Basins that fall above the upper bound are from the northwestern region where an increase in precipitation can reduce actual evapotranspiration from its climatic value (due to feedback effect), thereby producing increased runoff.

4. Hydroclimatology of the Continental United States

[8] Figure 2a illustrates the aridity index ϕ , defined as the ratio of mean annual potential evapotranspiration $\bar{P}\bar{E}$, to mean annual precipitation \bar{P} . Figure 2a illustrates that most

water resources regions (Blue lines in Figure 2 depict the boundaries of 18 water resource regions) in the U.S. are either temperate or humid, with the exception of the midwestern and southwestern regions. The maps depicted in Figures 1a and 2a–2c were created based on a fixed radius (0.5 degree) interpolation method within the ArcView geographic information system. Figure 2b illustrates the runoff ratio defined as the ratio of mean annual runoff \bar{Q} to mean annual precipitation \bar{P} . Comparing Figures 2a and 2b, arid regions correspond to regions with low runoff ratios. Figures 2a and 2b exhibit a marked east-west gradient indicating that as the aridity index decreases, the runoff ratio increases. The runoff ratio is very high in the northwest and northeast, since these regions are hilly with substantial cloud cover which in turn decreases incoming short-wave radiation and increases long-wave radiation from clouds, thereby decreasing the net radiation available at the earth's surface for evaporation. Figure 2c illustrates the runoff variability ratio, the ratio of standard deviation of annual streamflow σ_Q , to standard deviation of annual precipitation σ_P . Arid regions in the Midwest and southwest exhibit very low runoff variability ratios. Since these regions are always moisture limited, changes in precipitation tend to produce roughly equal changes in actual evapotranspiration resulting in negligible changes in runoff. In contrast, most humid basins experience high runoff variability ratios implying that evapotranspiration does not act to buffer runoff in these regions as it does in arid and semi-arid regions. Interestingly, humid basins in the northeast and northwest exhibit runoff variability ratios greater than 1 due to a feedback effect. In these regions, a positive precipitation anomaly due to increased cloud cover will further reduce the net radiation available at the surface resulting in a negative correlation between precipitation and evapotranspiration producing increased positive anomalous runoff. In summary, Figure 2 illustrates that much of the spatial variability in land surface fluxes may be predicted, approximately, by a watershed aridity index. Note that the annual streamflow and precipitation statistics have been calculated on the basis of water years (not calendar years) to avoid the influence of snow accumulation, storage and snowmelt processes.

[9] A common approach for illustrating the long-term hydroclimatology of a region (termed Budyko's framework)

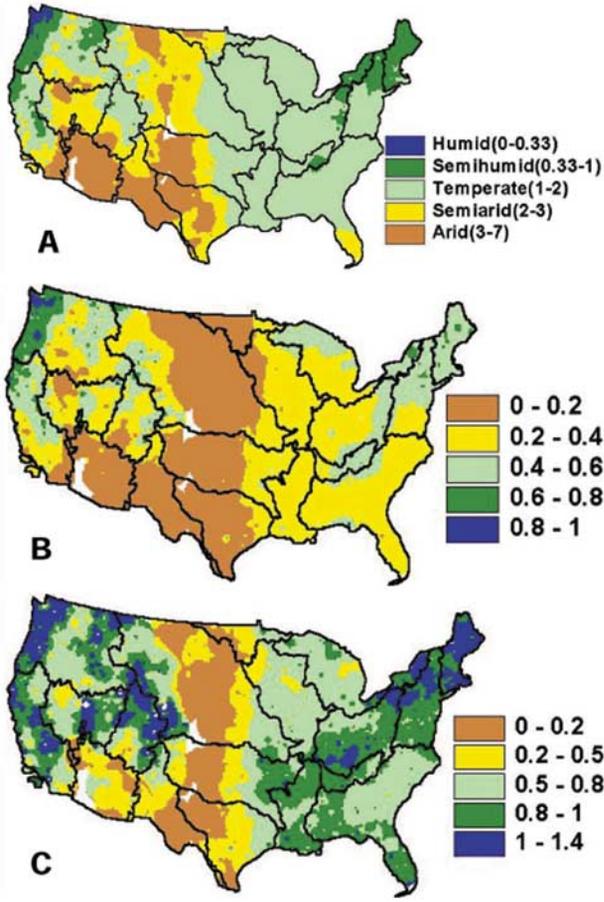


Figure 2. Hydroclimatological Indices for the U.S. (a) Aridity Index (b) Runoff ratio (c) Runoff variability ratio.

is to plot the evapotranspiration ratio \bar{E}/\bar{P} , estimated using $(\bar{P} - \bar{Q})/\bar{P}$, versus the basin aridity index $\phi = \bar{P}\bar{E}/\bar{P}$. Figure 3a illustrates one such relationship between \bar{E}/\bar{P} and ϕ introduced by Pike [1964]. Asymptotes A and B in Figure 3a describe the upper limits of the evapotranspiration ratio corresponding to minimum runoff potential. For humid regions with unlimited moisture supply, actual evapotranspiration approaches potential evapotranspiration which is illustrated by asymptote B in Figure 3a. Similarly, when potential evapotranspiration exceeds precipitation, then actual evapotranspiration approaches precipitation (Asymptote A in Figure 3a). Earlier studies [Budyko, 1974; Pike, 1964] suggested that the long-term water balance may be predicted by the aridity index alone. However, Figure 3a illustrates that the aridity index by itself, is unable to explain variations in the evapotranspiration ratio \bar{E}/\bar{P} .

[10] We take the approach introduced by Milly [1994] which exploits the structure of a water balance model to derive a theoretical relationship for \bar{E}/\bar{P} . Using the 'abcd' water balance model [Fernandez et al., 2000], we derived the following relationship for \bar{E}/\bar{P} as a function of the aridity index ϕ , and a new soil moisture index γ

$$\bar{E}/\bar{P} = 0.5 \left\{ 1 + \gamma(1 - R) - [1 - 2\gamma(1 - R) + \gamma^2(1 - 2R + R^2)]^{0.5} \right\} \quad (3)$$

where $R = \exp(-\phi/\gamma)$, $\phi = \bar{P}\bar{E}/\bar{P}$, $\gamma = b/\bar{P}$ and 'b' is the parameter of the 'abcd' water balance model. Figure 3b

illustrates the improved relationship in (3) for predicting the evapotranspiration ratio for three different values of γ . Figure 3b documents the importance of soil moisture storage, in addition to the aridity index, in determining the long-term water balance.

[11] Analogous to the approach taken by Budyko for the long-term water balance, Koster and Suarez [1999] attempted to predict the runoff variability ratio σ_Q/σ_P using the aridity index. Figure 4a compares the runoff variability ratio estimated using Koster and Suarez relationship with the observed runoff variability ratio at 1,337 watersheds in the continental U.S. Their relationship performs poorly for basins with high streamflow variability, again due to the omission of soil moisture in the runoff generation process. Basins with high soil moisture holding capacity will tend to reduce or buffer the variability in streamflow and vice versa. Similar to the previous results for the long-term water balance, accurate estimation of the runoff variability ratio requires a soil moisture index γ , in addition to an aridity index. A first-order approximation of the variance of streamflow predicted by the 'abcd' water balance model was used to derive the runoff variability relationship σ_Q/σ_P

$$\sigma_Q/\sigma_P = \left[\left(1 - \frac{\partial Y}{\partial W} \right) \left(1 - \frac{\partial S}{\partial Y} \frac{\partial Y}{\partial W} \right)^{-1} \right]_{\phi, \gamma, a} \quad (4)$$

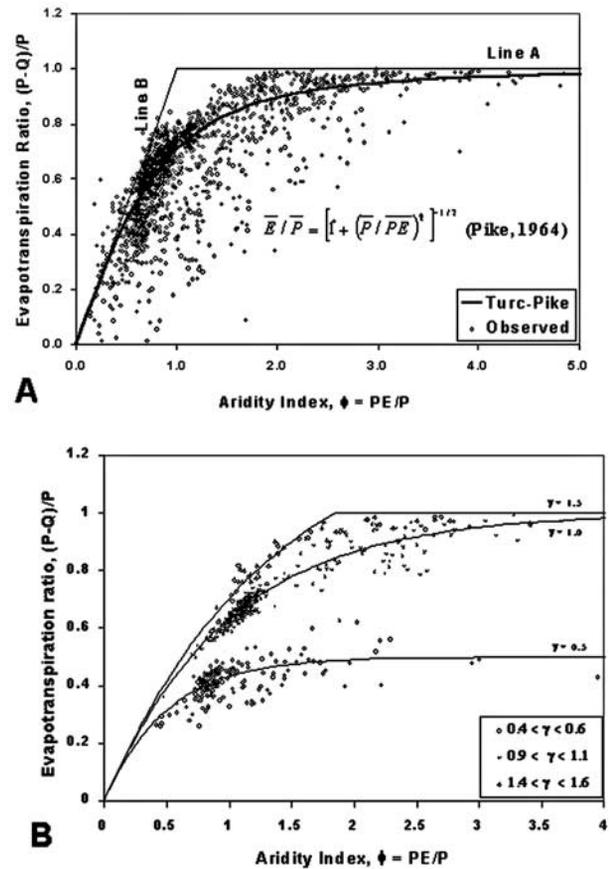


Figure 3. Comparison of Evapotranspiration ratio at 1,337 basins in the U.S., compared with (a) Turc-Pike relationship (b) Improved relationship (3) which depends on aridity index ϕ and soil moisture index γ .

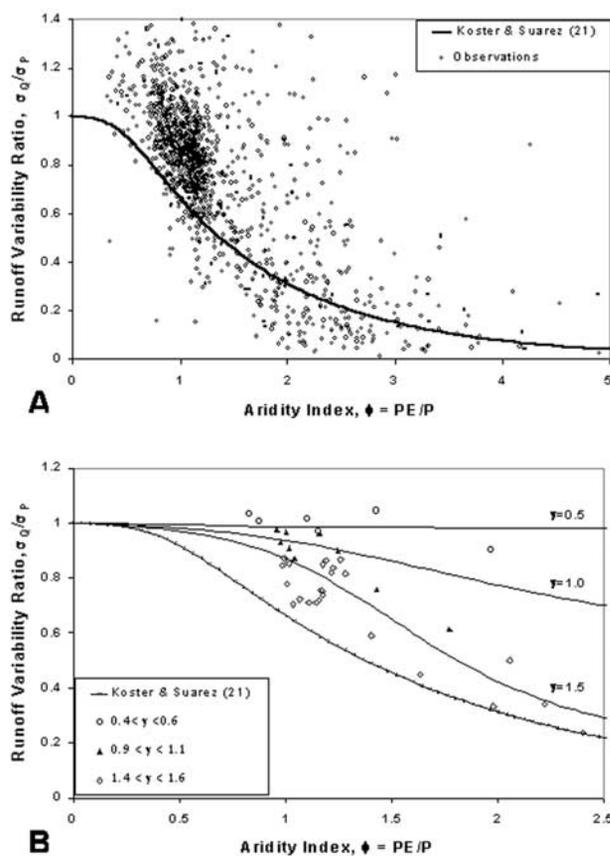


Figure 4. Interannual variability of streamflow in the U.S. (a) Prediction of runoff variability ratio using Koster and Suarez relationship (b) Enhanced curves for predicting runoff variability ratio (4) by incorporating the soil moisture index (γ). The slope of the fitted line between observed σ_Q and the σ_Q predicted by (4) is 1.031, whereas the slope of the fitted line between observed σ_Q and the σ_Q predicted by Koster and Suarez [1999] is only 0.71.

Figure 4b shows the enhanced relationship in (4) for predicting the runoff variability ratio for three different values of γ . Inclusion of both a soil moisture index and an aridity index led to gross improvements in our ability to predict the runoff variability ratio.

5. Concluding Remarks

[12] Our explorations of the long-term water balance, interannual variability of runoff and the sensitivity of streamflow to changes in precipitation provide an overview of the hydroclimatology of the continental United States. Using the concept of precipitation elasticity of streamflow, the sensitivity of streamflow to changes in precipitation is shown to be highly nonlinear with small changes in precipitation resulting in much larger changes in runoff for nearly all regions of the U.S. The upper bound of precipitation elasticity of streamflow - the inverse of runoff ratio - provides valuable information for climate change studies regarding the maximum expected % change in annual land surface response for 1% change in annual precipitation.

[13] We found that a watershed aridity index is not an adequate predictor, by itself, of either the long term water balance or the variability of annual runoff. Instead, both watershed aridity and watershed soil water holding capacity are necessary together, to predict either the long-term water balance and/or annual runoff variability. The relative importance of watershed soil moisture storage as a predictor of the mean and variance of annual watershed runoff may have important implications for our ability to model land-atmosphere interactions and for incorporating the impact of land surface processes in climate models. The approach taken here for modeling the long term water balance in terms of watershed aridity and soil moisture holding capacity may also be combined with a new approach to the regionalization of watershed model parameters introduced by Fernandez *et al.* [2000] to enable estimation of watershed model parameters at unaged sites.

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