Macroscale water fluxes
2. Water and energy supply control of their interannual variability

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Received 9 July 2001; revised 15 April 2002; accepted 15 April 2002; published 23 October 2002.

[1] Controls on interannual variations in water and energy balances of large river basins (10,000 km^2 and greater) are evaluated in the framework of the semiempirical relation $\overline{E}/\overline{P} = [1 + (\overline{R}/\overline{P})^{-\nu}]^{-1/\nu}$ in which and E, P, and R are basin mean values of annual evaporation, precipitation, and surface net radiation, respectively, expressed as equivalent evaporative water flux, overbars denote long-term means, and ν is a parameter. Precipitation is interpolated from gauges; evaporation is taken as the difference between precipitation and runoff, with the latter determined from basin discharge measurements and a simple storage-delay model; and radiation is based on a recent analysis in which 8 years of satellite observations were assimilated into radiative transfer models. Objective estimates of precipitation errors are considered; results suggest that past estimates of ν may have been biased by systematic errors in estimates of precipitation. Under the assumption that the semiempirical relation applies also to annual values, long-term mean observations are sufficient to predict the sensitivity of annual runoff to fluctuations in precipitation or net radiation. Additionally, an apparent sensitivity of runoff to precipitation can be inferred from the observations by linear regression. This apparent sensitivity is generally in good agreement with the predicted sensitivity. In particular, the apparent sensitivity increases with decreasing basin $\overline{R}/\overline{P}$; however, slightly excessive apparent sensitivity (relative to the prediction) is found in humid basins of the middle latitudes. This finding suggests a negative correlation between precipitation and net radiation: the increase in runoff caused by a positive precipitation anomaly is amplified by an accompanying decrease in surface net radiation, possibly induced by increased cloud cover. The inferred sensitivity of radiation (water flux equivalent) to precipitation is on the order of −0.1. Such a value is supported by independent direct analysis of annual precipitation and radiation data. The fraction of interannual variance in runoff explained by the annual precipitation anomaly (including any correlative influence of net radiation) varies systematically with climatic aridity, approaching unity in humid basins and falling to 40−80% in very arid basins. We conclude that the influence of seasonality of the precipitation anomaly on annual runoff is negligible under humid conditions, though it may be significant under arid conditions. *INDEX TERMS:* 1812 Hydrology: Drought; 1818 Hydrology: Evapotranspiration; 1854 Hydrology: Precipitation (3354); 1878 Hydrology: Water/energy interactions; 3359 Meteorology and Atmospheric Dynamics: Radiative processes; *KEYWORDS:* water balance, interannual variability, runoff, radiation

Citation: Milly, P. C. D., and K. A. Dunne, Macroscale water fluxes, 2, Water and energy supply control of their interannual variability, *Water Resour. Res.*, *38*(10), 1206, doi:10.1029/2001WR000760, 2002.

1. Introduction

1.1. Motivation

[2] When one speaks of a "wet year" or a "dry year," it is generally understood that such phrases refer to positive or negative anomalies of both precipitation and runoff, so strong is the correlation between these two fluxes at the annual timescale. The link between precipitation and runoff anomalies is common knowledge and underlies much of the work of surface water hydrologists. The generality of this rule motivates certain fundamental hydrologic questions of practical significance:

[3] If annual precipitation departs from its normal value by a unit depth, by how much will the annual runoff depart from its normal value? That is, what is the sensitivity of runoff to precipitation? Anomalous precipitation will be partitioned between runoff and evaporation. Thus it can be expected that the runoff sensitivity typically has an upper bound of 1, but may be much smaller under certain conditions. Considering the extreme case of a desert environment, we expect that the sensitivity is likely to be very small; it may even be zero. In contrast, a very humid river basin, in which all evaporative demand is normally met, could be expected to convert the entire precipitation anom-

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aly to runoff. Do observations support these hypotheses? If so, can we develop a simple rule to quantify this dependence of runoff sensitivity on climatic conditions?

[4] Does the seasonal distribution of the precipitation anomaly generally have a substantial effect on annual runoff? Empirical hydrologic analyses suggest that there is no unique answer to the first question, and that the runoff anomaly may also depend on the seasonal distribution of the precipitation anomaly [*Linsley et al.*, 1982, p. 256]. An anomaly of precipitation during the wet (or cold) season may produce more runoff than the same anomaly during the dry (or hot) season, because the latter is more likely to be consumed by evaporation. However, the importance of this effect will itself be a function of the seasonality of climate. For example, if the basin were wet throughout the year, we would expect to find that the runoff anomaly is uniquely determined by the total precipitation anomaly, regardless of its temporal distribution. Do the data support such speculation? Can we develop simple rules that suggest when the seasonality of the precipitation anomaly may be important, perhaps as a function of climate?

[5] To what extent do interannual variations in the amount of energy supplied to the land influence runoff? The predominant role of precipitation variability in forcing runoff variability probably tends to mask any possible influence of temporal variations in energy supply. Furthermore, hydrologists have generally been limited until recently to working with indirect measures of energy supply, such as air temperature and indices derived therefrom. The advent of new global observations of the Earth's radiation balance present an opportunity to begin to investigate the role of energy variations as a control on runoff.

[6] *Koster and Suarez* [1999] recently suggested a simple way to answer the first of these three questions by perturbative expansion of *Budyko*'s [1974] semiempirical water balance equation; similar approaches have been used to estimate the sensitivity of river basins to changes in mean climatic conditions [*Dooge*, 1992]. *Koster and Suarez* [1999] showed that this approach successfully described the runoff sensitivity of their numerical water balance model. Their analysis also implied that the roles of seasonal distribution of precipitation anomaly and interannual variability of energy supply (the subjects of the second and third questions above) were not great; however, their results apply only to their land model, embedded in an atmospheric general circulation model, and have not been confirmed by observational data.

1.2. Objectives of These Papers

[7] This is the second in a series of three papers analyzing controls on water balances of large land areas. Part 1 [*Milly and Dunne*, 2002] describes the development of the data set upon which the subsequent papers are based, with special attention to assessment of errors in estimates of precipitation. In the present paper (part 2), these data are employed to analyze the control of interannual water balance variations by fluctuations in supplies of water (precipitation) and energy (surface net radiation). In part 3 [*Milly and Wetherald*, 2002], the data of part 1 and the results of part 2 are used to develop and quantify a conceptual picture of land-process controls of monthly streamflow variability.

[8] Our objective here in part 2, more specifically, is to address the three questions posed in the Motivation section, through an interpretive analysis of observational data, within the framework established by *Budyko* [1974] and *Koster and Suarez* [1999]. We use long-term (more than 20 years) monthly time series of precipitation and discharge for a subset of the river basins analyzed in part 1. Objective measures of errors in precipitation estimates are used as the basis for selection of the basins. We supplement these data with monthly estimates of basin mean net radiation derived from fields supplied by the surface radiation budget (SRB) project of the Langley Research Center of the National Aeronautics and Space Administration. A secondary objective of this work is to begin to assess the value of such satellite-inferred radiation fields for hydrologic analysis.

[9] The questions posed above are phrased in terms of runoff. This focus is motivated by the observability, through stream discharge, of the runoff flux, and by the practical relevance of runoff for a wide variety of applied problems. An understanding of the control of runoff, however, translates more or less directly to an understanding of the control of evaporation, because evaporation and runoff anomalies sum to the precipitation anomaly. (The matter is complicated somewhat by interannual storage, which is treated here by a simple storage parameterization and by the use of water years for analysis.) Furthermore, sensible heat flux is the difference between net radiative energy supply and latent heat flux. Thus whatever controls evaporation variability will also be the dominant control, along with net radiation, of variability in sensible heat flux. In summary, then, an understanding of the control of runoff by water and energy supplies provides insight into the control of variability of all major land water and energy fluxes.

2. Data

[10] We use the river basin data set developed in part 1. The data set includes continuous, long-term (median record length 54 years) monthly time series of precipitation and discharge for 175 large (median area 51,000 km^2) basins worldwide. Discharge time series were produced by national hydrologic agencies using standard streamgauging techniques. The basin mean precipitation was estimated by interpolation of point gauge values. In addition, an 8-year, satellite-derived data set of monthly global fields of surface net radiation was used to produce an 8-year time series of basin net radiation.

[11] In part 1 we quantified the uncertainty in our estimates of basin mean precipitation. Statistical behavior of relative errors in the long-term annual mean were summarized by a parameter ψ_a,

$$\psi_a = \left[E\{\varepsilon_a^2\}\right]^{1/2}/\langle\hat{p}_a\rangle, \tag{1}$$

in which $E\{\ \}$ is the expectation operator, $\langle\hat{p}_a\rangle$ is the estimate of long-term basin mean annual precipitation, and ε_a is the error in that estimate. Components of ε_a considered include expected spatial-sampling errors in the absence of orographic effects, spatial-sampling errors associated with orographic effects, and errors in adjustments for gauge bias. We also applied standard correlation-based methods to develop estimates of the standard errors of anomalies from the mean during any particular month or year. As a summary measure of the anomaly errors, we introduced the parameter $\overline{\psi}_n$,

$$\overline{\psi}_n = \overline{\sigma_n^2}/Var(\langle P_n \rangle), \quad (2)$$

in which σ_n^2 is the variance of the error in the estimate of the basin mean anomaly for year, an overbar denotes an average over the period of record, and $Var(\langle P_n \rangle)$ is the variance of the basin-mean annual precipitation.

[12] For this analysis, we first chose the subset of the 175 basins of part 1 for which ψ_a was less than 0.1 and area of lakes was no more than 25% of basin area. These conditions yielded 73 basins for the first section of our data analysis, in which we limit our attention to the average annual water balance. In the second part of the data analysis, we examine interannual variability. For this purpose, we further restricted the set of basins to those 42 basins for which $\overline{\psi}_n$ was less than 0.05.

[13] Our analyses of variability are performed with annual mean variables defined on a water year basis. We define a unique water year a priori for each basin. The water year is defined as the 12-month period beginning with month m, where m is chosen from the set (1, 2, ..., 12) in order to maximize the explanation of variance in a simple linear regression of discharge against same-year precipitation for the full period of record of each basin. We used water years instead of calendar years to minimize the errors associated with our simple treatment of carryover storage.

3. Theory

3.1. Defining Runoff Sensitivities to Precipitation β and Net Radiation γ

[14] We first introduce a definition of the sensitivities of runoff to precipitation and net radiation. Subsequently, we shall use a semiempirical water balance relation to develop predictors of these sensitivities. The choice of precipitation and net radiation as independent variables is motivated by the form of the semiempirical relation. We distinguish conceptually between runoff and discharge. Runoff is produced by the interaction of land water and energy supplies, mainly through storage in the root zone of the soil. Runoff differs from discharge in that the latter is lagged in time as a result of storage "downstream" from the runoff-producing region (*e.g.*, in aquifers, stream channels, and riparian subsurface storage zones). Let Q, P, and R denote basin-mean values of observational estimates of annual amounts of runoff, precipitation, and surface net radiation, respectively. We shall express radiation in units of equivalent evaporative water fluxes, so that Q, P, and R, can all be expressed in a common set of units. (Although our notation here is generally consistent with that of part 1, we henceforth discontinue use of the "hat" to denote observation-based estimates of a quantity and the angle brackets to denote basin mean values.) We assume that the expected runoff anomaly (δQ_n) during any year n has a component proportional to the precipitation anomaly (δP_n) during the same period, a component proportional to the net radiation anomaly (δR_n), and a zero mean random component (u_n),

$$\delta Q_n = \beta \delta P_n + \gamma \delta R_n + u_n. \quad (3)$$

The random component is at least partly associated with variations in the spatial and (intraperiod) temporal distributions of precipitation and net radiation, and also with interperiod variations in any other climatic or surface controls on runoff. We shall call β and γ the sensitivities of runoff to precipitation and net radiation, respectively.

3.2. Sensitivity Estimates from Data

[15] Despite its simple form, (3) cannot be used directly for data analysis, because we have observations only of discharge, not of runoff. Additionally, long time series of radiation estimates generally are not available. Because radiation may be correlated with precipitation, even estimates of β made from analysis of precipitation and runoff time series would be biased by the neglect of radiation. In this section, we present our strategy for dealing with these problems.

[16] To relate the runoff anomaly to the (observable) discharge anomaly (δY_n), we assume

$$\delta Y_n = (1 - \alpha)\delta Q_n + \alpha \delta Y_{n-1} + w_n, \quad (4)$$

in which α is a constant and w_n is another error term; this equation produces a lag between runoff and discharge and forces the two fluxes to balance over time. The parameter α is a measure of the persistence of the annual streamflow series. Physically, α parameterizes the importance of interannual storage in the basin. Combination of (3) and (4) yields

$$\delta Y_n = (1 - \alpha)\beta \delta P_n + (1 - \alpha)\gamma \delta R_n + \alpha \delta Y_{n-1} + w_n \quad (5)$$

in which w_n has been redefined as a linear combination of u_n and the original w_n, which are henceforth abandoned.

[17] The other issue is that of availability of radiation data. In preliminary analyses, we explored the utility of (5) for determination of the sensitivities by multiple regression analysis. Because estimates of R_n are available only for a limited time span (8 years, or generally 7 complete water years), standard errors of the estimated coefficients were unacceptably large. To mitigate this problem, we introduce a second linear relation between precipitation and radiation,

$$\delta R_n = \rho \delta P_n + \delta r_n, \quad (6)$$

in which δr_n represents the part of the anomaly δR_n that is not correlated with δP_n. Combination of (5) and (6) yields

$$\delta Y_n = (1 - \alpha)(\beta + \gamma\rho)\delta P_n + \alpha \delta Y_{n-1} + w_n \quad (7)$$

in which we have lumped the unknown δr_n into a redefined error w_n. We use (7) to estimate α and $\beta + \gamma\rho$ by regression from discharge and precipitation time series. The quantity $\beta + \gamma\rho$ is an apparent sensitivity of runoff to precipitation, which includes the indirect effect on runoff associated with the induced changes in energy supply. The parameter ρ can be estimated independently through (6), albeit with considerable uncertainty, by regression over the shorter period of availability of radiation data. The values of γ and, hence, β are still undetermined, however, requiring that we

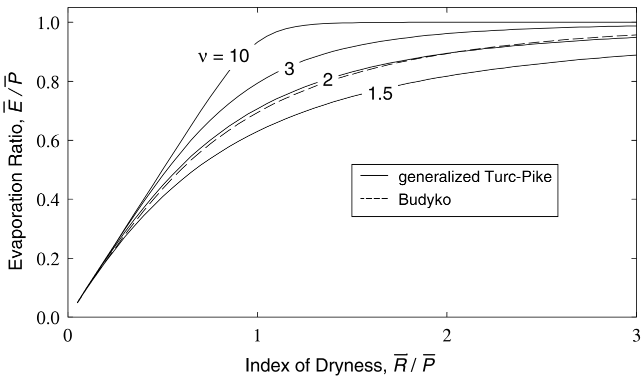

Figure 1. Relations between evaporation ratio, $\overline{E}/\overline{P}$, and index of dryness, $\overline{R}/\overline{P}$, assumed in semiempirical water balance equations of the form (8). The generalized Turc-Pike function is given by (10), and the Budyko function is given by (9).

introduce additional theoretical information in order to constrain our results further.

3.3. Estimating β and γ from Semiempirical Water Balance Theory

[18] *Budyko* [1974] presented a semiempirical expression for average water balance partitioning as a function of the relative magnitudes of water and energy supply rates,

$$\overline{E}/\overline{P} = \phi(\overline{R}/\overline{P}), \tag{8}$$

in which E is evaporation, an overbar denotes an average over the period of record, and ϕ is a "universal" interpolation function, assumed applicable for any river basin. Budyko's approach is termed semiempirical because the interpolation function was determined empirically as a best fit to data, but the dimensionless argument (the index of dryness, $\overline{R}/\overline{P}$) is determined by dimensional analysis and the form of the interpolation function was constrained to obey hypothesized physical constraints that $\overline{E}$ approach $\overline{P}$ in the limit of arid climates (high index of dryness) and that $\overline{E}$ approach $\overline{R}$ in the limit of humid climates (low index of dryness). *Budyko* [1974] chose

$$\phi(x) = \phi_B(x) \equiv \left[x\left(\tanh\left(x^{-1}\right)\right)(1 - \cosh x + \sinh x)\right]^{1/2} \tag{9}$$

Budyko's most significant contribution to (8) was not so much the empirical form of $\phi_B(x)$ as it was the identification of net radiation as a control on evaporation under conditions of ample water supply. Indeed, many relations akin to (8) had already been proposed [*Brutsaert*, 1982]. In particular, one useful family of functions that can be used to describe $\phi(x)$ is [*Choudhury*, 1999]

$$\phi(x) = \phi_\nu(x) = \left[1 + x^{-\nu}\right]^{-1/\nu}, \tag{10}$$

where ν is a curve parameter. We refer to this as the generalized Turc-Pike relation; for the parameter value ν = 2, it reduces to the Turc-Pike relation [*Turc*, 1954; *Pike*, 1964]. With ν = 2, (10) differs little from (8); as ν becomes large, the curve approaches its asymptotes for all levels of aridity (Figure 1).

[19] If (8) is an adequate description of relations among long-term (multidecadal) average fluxes, then it might also be adequate to describe relations among fluxes at shorter timescales [*Koster and Suarez*, 1999]. Indeed, although we have introduced the relations above in terms of long-term means, in many cases they were actually developed to describe interannual variability within a given basin [*Brutsaert*, 1982]. Strictly speaking, a single function $\phi(x)$ cannot apply both to annual ratios and to long-term ratios, as a result of nonlinearity. In practice, the interannual variability is typically small enough that this complication can be ignored. We shall assume that different basins can have different values of ν, but that interannual variability for any basin can be characterized by a time-invariant value of ν. If we assume also that departures of fluxes from their long-term mean are sufficiently small, perturbation of (8) implies

$$\delta E_n = \left[\phi - \left(\overline{R}/\overline{P}\right)\phi'\right]\delta P_n + \phi'\delta R_n, \tag{11}$$

in which ϕ' is the derivative of ϕ with respect to its argument. Because Q is the difference between P and E, we also have

$$\delta Q_n = \left[1 - \phi + \left(\overline{R}/\overline{P}\right)\phi'\right]\delta P_n - \phi'\delta R_n. \tag{12}$$

By comparison with (3), this relation provides expressions for the runoff sensitivities to precipitation and radiation,

$$\beta = 1 - \phi + \left(\overline{R}/\overline{P}\right)\phi', \tag{13}$$

$$\gamma = -\phi'. \tag{14}$$

Adopting the generalized Turc-Pike equation (10) in conjunction with (8), we find that these sensitivities are functions of the index of dryness and the parameter ν. (And for Budyko's $\phi(x)$, they depend only on the index of dryness.) Because the evaporation ratio itself is also a function of the index of dryness and ν, the index of dryness can be eliminated to arrive at relations between the evaporation ratio and the runoff sensitivities β and γ. Magnitudes of β (Figure 2) and γ (Figure 3) approach zero under very arid conditions and unity under very humid conditions. The magnitude of β grows

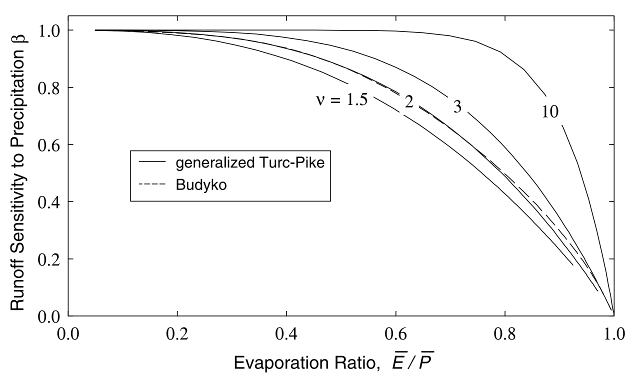

Figure 2. Runoff sensitivity to precipitation β as a function of evaporation ratio, as predicted by the generalized Turc-Pike relation (10), for selected values of the parameter ν. Sensitivity based on Budyko's relation is shown for comparison.

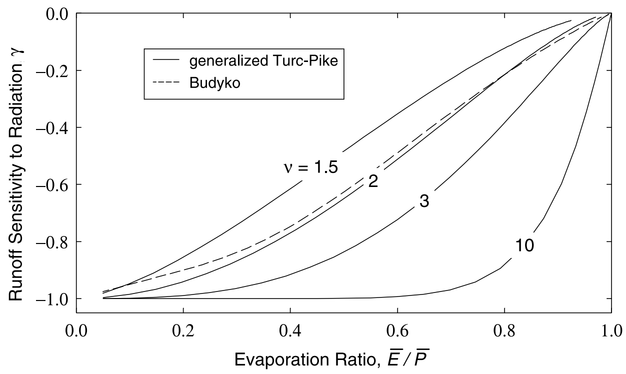

Figure 3. Runoff sensitivity to surface net radiation γ as a function of evaporation ratio, as predicted by the generalized Turc-Pike relation (10), for selected values of the parameter ν. Sensitivity based on Budyko's relation is also shown for comparison.

more rapidly with decreasing evaporation ratio than does that of γ, which implies that water balance is more sensitive to precipitation than to radiation at intermediate values of the evaporation ratio.

4. Data Analysis

4.1. Average Annual Water Balance

[20] We display our water balance data, in the framework of (8), in Figure 4. For most basins, evaporation is greater than that predicted by the Turc-Pike equation ($\nu = 2$) [*Turc*, 1954; *Pike*, 1964] or by the similar equation of *Budyko* [1974] (not shown), but does not exceed limits associated with water or energy supplies (the "supply asymptotes"). In many of the basins, the departure from $\nu = 2$ is larger than could be explained reasonably by estimated precipitation error, as indicated by the error lines in Figure 4. (Recall that only basins with characteristic relative precipitation error ψ_a smaller than 0.1 were included in the analysis.) The most significant exceptions to the general tendency for points to lie between $\nu = 2$ and the supply asymptotes are several humid basins (index of dryness less than 1) with unexpectedly low evaporation (i.e., $\nu < 2$), and a few humid basins with evaporation greater than the equivalent radiative energy availability (i.e., $\overline{E}$ greater than $\overline{R}$).

[21] The tendency for evaporation to exceed that predicted by the Turc-Pike and Budyko relations implies a systematic difference between our analysis and previous analyses. This could be explained by a systematic difference in estimates of precipitation, discharge, and/or energy supply (net radiation or potential evaporation). In particular, the discrepancy might result from insufficient attention to precipitation bias in previous analyses. To explore this idea, we replotted Figure 4, using the entire data set of part 1, with no restriction on precipitation error (Figure 5). Most of the additional basins mainly plot as points below the $\nu = 2$ curve as a result of the strong tendency of the largest errors in precipitation to be negative; negative errors arise from gauge undercatch, especially in regions of substantial snowfall, and from spatial sampling errors in regions of high topographic relief. We fitted (10) to our data by minimizing the sum of the squared deviations of predicted and observed evaporation ratios. We found $\nu = 1.5$ when all data were used; $\nu = 2.0$ when only basins with $\psi_a < 0.34$ (basins with obviously biased precipitation errors excluded) were used; and $\nu = 2.55$ for $\psi_a < 0.1$ (i.e., for the data set used to generate Figure 4). Thus our findings would be consistent with those of previous studies ($\nu = 2$) if we had exercised only moderate selectivity with respect to precipitation errors.

[22] An additional factor that, in part, may explain departures from classical curves in some arid basins is human disturbance of natural water fluxes. Development and consumptive use of water resources in arid basins increases total evaporation at the expense of runoff. We reviewed available water use data and concluded that this

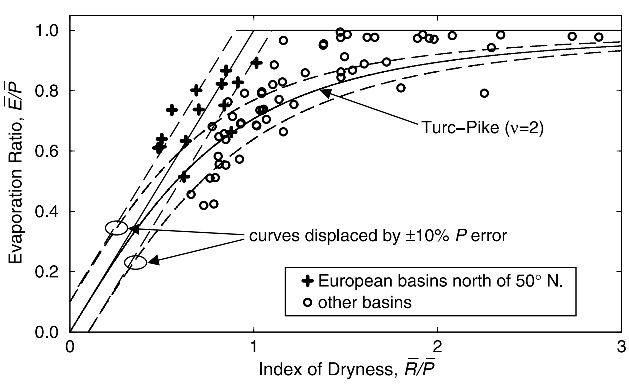

Figure 4. Scatterplot of evaporation ratio, $\overline{E}/\overline{P}$, against index of dryness, $\overline{R}/\overline{P}$. Each symbol represents one basin. Evaporation is computed as the difference between precipitation and runoff. Solid curves are the Turc-Pike relation ($\nu = 2$) and the asymptotes; dashed lines show the displacements from them that would be induced by positive and negative precipitation biases of 10%. Crosses mark basins whose centroids lie north of 50°N.

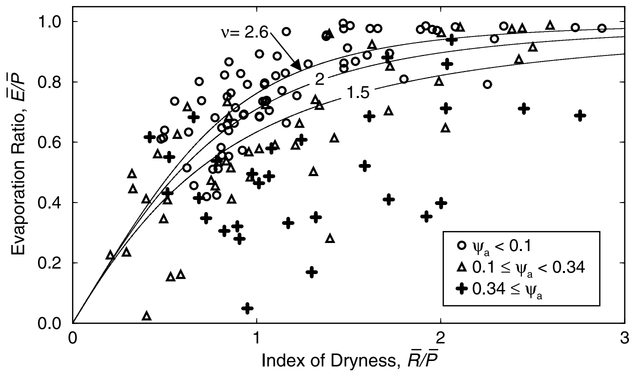

Figure 5. Scatterplot of evaporation ratio, $\overline{E}/\overline{P}$, against index of dryness, $\overline{R}/\overline{P}$, for the 155 basins of part 1 for which at least 5 gauges were used in the precipitation analysis and no more than 25% of basin area is covered by lakes. (Estimates of ψ_a are valid only when at least 5 gauges are available.) Also shown are fitted generalized Turc-Pike equations that minimize squared deviations for these 155 basins ($\nu = 1.5$), for the 122 such basins for which ψ_a is also less than 0.34 ($\nu = 2$, the Turc-Pike value), and for the 73 such basins for which ψ_a is also less than 0.1 ($\nu = 2.6$). (Data points for seven basins are beyond the boundaries of the plot but were included in the calculation of best fit curves.)

factor could be important only in a small fraction of the arid basins; disturbance generally was not of sufficient magnitude to explain a large part of the discrepancy between observations and the usual curves.

[23] The foregoing discussion can explain evaporation in excess of that predicted by the Turc-Pike and Budyko expressions, but cannot explain why evaporation could exceed that given by the energy supply asymptote ($\overline{E} = \overline{R}$). A positive bias in precipitation estimates, averaging about 10%, would be sufficient to explain the location of these points above the energy supply asymptote in Figure 4. Given our basin-selection criteria, such a bias seems possible for such a small fraction of the overall set of basins; however, it is notable that all of the basins in which evaporation exceeds the energy limit are located in northern Europe, and that evaporation in all basins in northern Europe is near or greater than the energy limit. It is unlikely that such similar precipitation errors would arise by chance. One possible explanation is that evaporation in these basins is assisted by a second, unaccounted (i.e., other than radiation) energy source. Oceanic air masses, warmed by the North Atlantic Current (Gulf Stream), may deliver substantial additional energy by large-scale advection of sensible heat. Sensible heat flux to the surface is common in high-latitude winters; explanation of our results would require a net downward sensible heat flux even in the annual mean. It is also possible that the radiative algorithms used to estimate $\overline{R}$ have a substantial negative bias specific to this region, perhaps associated with regional cloud, humidity, and/or temperature patterns induced by the North Atlantic Current.

4.2. Interannual Variability

[24] For the analysis of interannual variability, the data set was reduced to include only those basins for which variance of the error in our estimate of the annual precipitation anomaly was smaller than 5% of the variance of the annual precipitation itself ($\overline{\psi}_n < 0.05$). As noted in the Data section, this criterion reduces the size of our data set from 73 to 42 basins. Using long-term precipitation and discharge

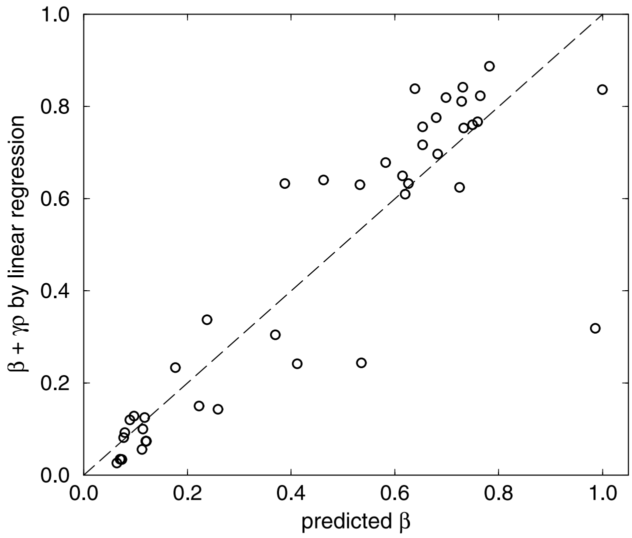

Figure 6. Scatterplot of apparent runoff sensitivity to precipitation, $\beta + \gamma\rho$, determined by regression of historical annual data, against runoff sensitivity β predicted by (13).

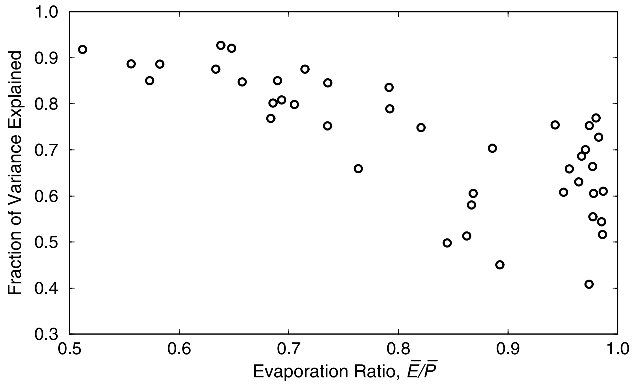

Figure 7. Scatterplot of fraction of variance in annual runoff anomalies explained by annual precipitation anomalies, according to (7), against evaporation ratio, $\overline{E}/\overline{P}$.

records, we estimated α and $\beta + \gamma\rho$ for each basin by least squares regression applied to (7). For most basins, the estimated value of α was between 0 and 0.3, indicating a certain degree of persistence in annual flows due to interannual basin storage. Values of α were slightly negative for four basins, presumably because of random noise in the estimation process. Estimated values of $\beta + \gamma\rho$ ranged from 0.027 to 0.89 and were distributed bimodally; values were small for dry basins and large for humid basins, as expected qualitatively for β alone from Figure 2.

[25] For comparison, we also estimated β by means of the generalized Turc-Pike equation. For each basin, the observation-based estimates of long-term evaporation ratio and radiative index of dryness were used with (8) and (10) to determine a value of ν. Then β was estimated from (13). For those few basins in which evaporation appears to exceed its energy-limited value, ν cannot be defined; in these cases, we simply set ν to a large value (implying that ϕ coincides with the supply asymptotes).

[26] In Figure 6, we compare the theoretical runoff sensitivity estimated from the long-term means through (8) with the apparent sensitivity determined by regression. Perfect agreement would be expected if all data were error-free, if interannual variability were characterized by a constant value of ν for any given basin, and if ρ (correlation between precipitation and radiation variations) were zero (or radiation did not vary interannually). In general, the agreement is very good, implying that any radiative effects are subordinate to direct control by precipitation; a detailed analysis is presented below. Two notable outliers are associated with apparent sensitivities much smaller than those predicted by (8). These both represent basins in northern Europe, for which available energy seems to have been significantly underestimated, as already discussed; the other basins in this region are not represented in Figure 6 because the magnitudes of their precipitation anomaly estimation errors were too large for inclusion in the analysis of variability.

[27] Having addressed the first question presented in the Introduction, we now turn to the second. We saw in Figure 6 that the apparent sensitivity of annual runoff to annual precipitation is generally consistent with expectations from the semiempirical model. Figure 7 shows how much of the total variance of annual runoff is explained by this sensitivity.

Eighty to ninety percent of the annual discharge variance for basins in humid climates (evaporation ratio less than 0.7) is explained by annual precipitation. This finding leaves little room for additional influences, consistent with the hypothesis, advanced in the Introduction, that runoff anomalies of humid basins should be independent of the seasonal distribution of the precipitation anomaly. Even in arid climates (evaporation ratio greater than 0.8), the annual total precipitation anomaly generally explains more than half of the runoff variability. A substantial part (i.e., one-quarter to one-half) of the variance, however, can be attributed to other factors in these cases. Presumably, the seasonal distribution of the precipitation anomaly is a major factor.

[28] Aside from the overall agreement and the two outliers in Figure 6, the most remarkable feature of Figure 6 is the tendency for symbols representing large values of sensitivity (corresponding to humid conditions) to lie above the 1:1 line . Here we explore the possibility that this feature may be explained by positive values of the product $\gamma\rho$. From (14) and the general expectation that ϕ is a monotonically increasing function, we expect γ to be negative. Positive values of $\gamma\rho$ would then imply negative values of ρ. The physical interpretation would be that positive fluctuations in precipitation tend to produce proportionate negative fluctuations in surface net radiation. These fluctuations, in turn, amplify the positive anomaly in runoff. Consideration of energy supply variability brings us to the subject of the third question in the Introduction. Let us use β_ν to denote the value of β predicted from the semiempirical relations and the long-term mean observations, and let $(\beta + \gamma\rho)_\delta$ denote the apparent sensitivity determined by linear regression of annual anomalies of precipitation and discharge. If we assume equivalence of β_ν and the β in $(\beta + \gamma\rho)_\delta$, and if we assume also that (14) can be used, together with the generalized Turc-Pike model, to estimate γ (with the estimate denoted γ_ν), then we can estimate ρ by

$$\rho = \frac{(\beta + \gamma\rho)_\delta - \beta_\nu}{\gamma_\nu}. \qquad (15)$$

Values of ρ so estimated are plotted against evaporation ratio in Figure 8. Inferred values scatter widely under arid

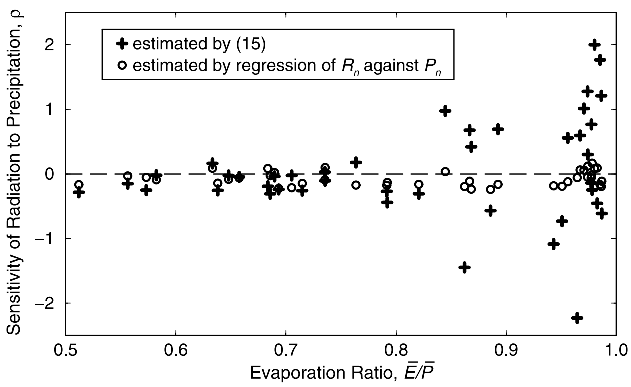

Figure 8. Scatterplot of inferred values of radiation sensitivity to precipitation ρ against evaporation ratio, $\overline{E}/\overline{P}$. Crosses represent values obtained indirectly by (15); circles represent values obtained by direct regression of SRB data against precipitation. Dashed line shows $\rho = 0$.

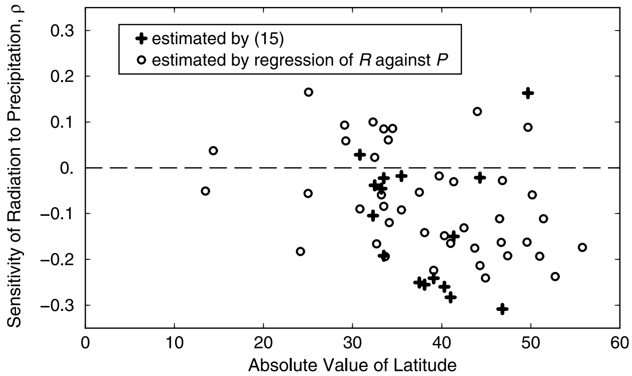

Figure 9. Scatterplot of estimates of radiation sensitivity to precipitation ρ against absolute value of latitude of the basin centroid. Crosses represent values obtained indirectly by (15), circles represent values obtained by direct regression of SRB data against precipitation, and dashed line shows $\rho = 0$. For estimates based on (15), only results for evaporation ratios smaller than 0.75 have been plotted, because of the large scatter noted in Figure 8 for basins with larger evaporation ratios.

conditions (evaporation ratio approaching 1), but are relatively stable and generally negative for lower values of the evaporation ratio. The instability of ρ under dry conditions can be understood by reference to (15). Under dry conditions, the denominator and both terms in the numerator are expected to approach zero. Small errors in estimates of γ_ν then lead to wide scatter of the ρ estimates. Also, small errors in the terms in the numerator allow it to take either sign, regardless of its true value.

[29] In contrast, we expect such estimation errors to be minimized (at least in a relative sense) for lower values of the evaporation ratio. Figure 8 shows estimated values of ρ; most values are in the range from -0.3 to 0, with a mean of -0.12, for evaporation ratio less than 0.75. For comparison, we estimated ρ in (6) by regression, using the SRB data. Inferred values tend to be negative and small, as seen in Figure 8, as do the more reliable (more humid basin) values derived from (15). Figure 9 shows both estimates of ρ as a function of latitude. Standard errors of these ρ estimates (not plotted for clarity) are generally large, and few of the ρ values are significantly different from zero. Overall significance can be judged subjectively by looking at the data collectively. Both approaches to estimation of ρ suggest a negative value of ρ for middle and high latitudes; estimates are closer to zero for latitudes within about 35° of the Equator (but data for the tropics are scarce).

[30] A detailed analysis of the physical mechanisms responsible for a correlation between precipitation and surface net radiation is beyond the scope of this paper. In general, the increased cloud cover that accompanies a positive precipitation anomaly [e.g., *Plantico et al.*, 1990] is expected to reduce the amount of shortwave radiation reaching the surface, but to increase the net downward longwave radiation. Variations in atmospheric humidity may also affect the longwave radiation balance. Our review of the available empirical relations [*Brutsaert*, 1982] suggests that the decrease of the shortwave radiation flux should dominate the response of annual mean net radiation to changes in cloud cover, consistent with the inference here

of a negative correlation between precipitation and surface net radiation.

5. Summary and Discussion

5.1. Summary

[31] We evaluated controls on interannual variations in observed river discharge. We found that the sensitivity of runoff to precipitation can be predicted well from information on long-term mean water and energy balance. In humid basins of the middle latitudes, however, observational data imply a slightly greater sensitivity of runoff to precipitation than expected. This result is suggestive of a negative correlation between precipitation and net radiation: the increase in runoff caused by a positive precipitation anomaly is amplified by an accompanying decrease in surface net radiation, possibly induced by increased cloud cover. The inferred sensitivity of (water equivalent) radiative flux to precipitation is small and negative, on the order of -0.1. Such a value is supported by independent direct analysis of precipitation and radiation data.

[32] The fraction of interannual variance in runoff explained by the annual precipitation anomaly varies systematically with climatic aridity, approaching unity in humid climates, and falling as low as 40–80% in very arid basins. We infer that the influence of seasonality of the precipitation anomaly on runoff is negligible under humid conditions, but potentially important under arid conditions.

5.2. Observational Confirmation of Results of *Koster and Suarez* [1999]

[33] *Koster and Suarez* [1999] proposed a simple but powerful framework for analysis of interannual variability of land water and energy fluxes. Their approach rested on the hypothesis that interannual storage and interannual radiation anomalies could be ignored. This assumption leads to an ability to predict sensitivities from only long-term mean data. Koster and Suarez confirmed the effectiveness of their approach for analysis of fluxes in an atmospheric general circulation model. Here we have substantially confirmed the validity of this approach using observational data. It must be recalled, however, that we accounted for interannual storage in the analysis of discharge variations; we have found its effect to be important in some basins. Additionally, our analysis suggests that radiative flux variations may be detectable in variations of surface water fluxes.

5.3. Semiempirical Balance Relations and Precipitation Bias

[34] An interesting feature of our results is the tendency for observationally inferred evaporation ratios to be greater than those predicted by standard semiempirical water balance equations, such as that of *Budyko* [1974]. It is possible that our stringent basin-selection criteria are more selective than those used in previous studies and could explain the discrepancy. If basins with a large negative bias in precipitation estimates were used to fit the earlier relations, then those relations would tend to underpredict the evaporation ratio. Negative biases are common because of gauge undercatch, especially in cold regions, and because of spatial sampling errors in regions affected by orographic precipitation.

5.4. Inferences About the SRB Data Set

[35] This analysis provides an early application of recently available satellite-based estimates of surface net radiation to land water and energy balance problems. It is not possible to make conclusive inferences about the quality of the SRB data set from our analysis, but some points are worthy of mention. Overall, the SRB-based analysis of long-term mean water balances yielded results generally consistent with those of previous analyses; however, evaporation in northern Europe appeared to exceed the (water equivalent) net radiative energy supply. This finding is inconsistent with *Budyko*'s [1974] hypothesis that radiative energy supply is an upper bound for latent heat flux. Unless we have underestimated precipitation errors in this region, these results indicate either a departure from Budyko's hypothesis or a bias in the SRB radiation estimates.

[36] Our analysis also suggests, albeit indirectly, that the SRB data set may be capturing at least some of the interannual variability of net radiation. As mentioned above, a negative correlation between precipitation and SRB radiation is supported by the independent analysis of apparent runoff sensitivity to precipitation. While these results are encouraging, the negative correlation is generally weak. A longer radiation record would allow us to develop more accurate basin-specific estimates of radiation-precipitation sensitivity and to examine more directly the sensitivity of water and energy balances to radiative fluxes.

5.5. Practical Implications

[37] The most robust finding of our analysis is that semiempirical relations for long-term annual water balance can be used to derive, very easily, first-order estimates of the sensitivity of annual runoff and evaporation to annual precipitation. These sensitivities are uniquely determined by the ratio of long-term annual mean runoff to long-term annual mean precipitation. Sensitivities so derived would be most useful in humid climates, where annual precipitation has been shown to be a strong control of annual water fluxes. Possible applications of these sensitivities include drought analysis, analysis of hydrologic response to climate anomalies, and estimation of water resource sensitivity to climate change.

[38] **Acknowledgments.** Stephen A. Klein, Richard T. Wetherald, and two anonymous reviewers kindly reviewed a draft of the manuscript for us.

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